GEOLOGICAL MAGAZINE

VOL. XCV. No. 3. MAY–JUNE, 1958

Issued 3rd June, 1958

A Metamorphic Boundary in the Moine Schists of the Morar and Knoydart Districts of Inverness-shire

By R. St. J. Lambert

Abstract

A re-examination of the Moine—Sub-Moine boundary of the Morar dome shows it to be a zone of transition between two metamorphic states. Some notes on the structure accompany a mineralogical description of the changes across the boundary; discussion is made of evidence that the metamorphic boundary transgresses the geological boundaries. The schists of the core are thought to be in a lower metamorphic state than those of the envelope.

Introduction and Notes on Previous Research

THE area under consideration is the same as that described by Richey and Kennedy (1939), covering a small section of Knoydart and the larger part of Morar. These form part of the One-inch Sheet 61, which Richey, Kennedy, and others were mapping until the outbreak of war interrupted their work.

In Richey and Kennedy (1939) the schists of the area were divided into the Moine Series of Morar and the Sub-Moine Series, both including psammitic and pelitic schists. The Sub-Moine Series contained a group of hornblendic and acid gneisses which were comparable directly with those of Glenelg and Skye (notably Knock Bay) although the marbles, eclogites, and other extreme rock types of Glenelg have not been discovered at Morar. The Moine Series occupied the envelope of an anticline (or dome) while the Sub-Moine occupied the core. Supposedly distinctive differences between the two series led the authors to consider that the Sub-Moine Series was older than the Moine Series, having existed as schists before the deposition of the latter. It was observed that the Sub-Moine Series showed signs of retrograde metamorphism, but no detailed evidence was given.

It was claimed that the two series are separated by an unconformity, and Kennedy described recumbent folding within the core. Little remark was made upon the Hornblende Gneiss Group. The succession

is to be found in Table 1, where it is compared with MacGregor's version, Kennedy's revised version, and the present writer's views.

A. G. MacGregor, in taking up the preparation of the memoir and map after the war, commenced work on the south-east of the area, where the Moine–Sub-Moine boundary is least clearly defined. He was not impressed by the evidence provided by Richey and Kennedy and in 1948 published a review of the supposed differences between the two series. He discovered another type of calc-silicate rib in members of both series; he showed that a sutured quartz mosaic was not a necessary or diagnostic feature of the Sub-Moine Series, that foliation of the micas oblique to the bedding occurred locally in the Moine Series and that iron-rich epidote was not confined to the Sub-Moine. He reiterated and developed these conclusions in 1952.

In 1948 Richey described some complex drag folds in the Moine Series on the west of the anticline, concluding that the folding occurred before the regional metamorphism and crystallization of the present mineral assemblage.

The work of Kennedy on the calc-silicate bands (1949) demonstrated an increase in metamorphic grade across the district from west to east, towards the great belt of injection metamorphism. The limited distribution of these bands and more particularly their absence in the Sub-Moine Series makes this work difficult to apply in detail.

Kennedy's recent discussion of the structure (1955) combined a reconsideration of the results of the Morar-Knoydart work with that of studies in Skye (Clough, 1907; Bailey, 1939) and Glenelg (Clough, 1910) to produce a unified hypothesis. In this study the following points were assumed; that the orthogneisses (the hornblendic and acid gneisses) were an older series invariably acting as a base to the Moine Series, that the complex folding in the core of the anticline did not extend into the Moine Series and that the lower Psammitic Group (m′) represented the lowest member of the Moine Series. Reasons were advanced for each of these assumptions. Kennedy stated that the schists and gneisses of the core were autochthonous but highly folded, sliced, and thrust and that the schists of the envelope had been thrust over them, having been torn from their roots to form the Morar Nappe. This interpretation met with considerable opposition (see discussion in Kennedy, 1955).

The unsatisfactory nature of the present position with regard to the structure and metamorphic history is readily apparent. The divergence of published opinion is most acute on the status of the hornblendic group of the core, while a second conflict concerns the Moine-Sub-Moine boundary, which may be of stratigraphic (Richey and Kennedy, 1939), tectonic (Kennedy, 1955), or metamorphic origin (G. Wilson and others in discussion in Kennedy, 1955). The term " core-envelope

TABLE I

RICHEY AND KENNEDY 1939		MACGREGOR 1948	KENNEDY 1955		LAMBERT 1958	
Upper Psammitic Group m^3	MOINE Stratigraphic succession	MOINE with associated orthogneisses.	Upper Psammitic Group m^3	MOINE (envelope)	Upper Psammitic Group mE	MOINE (envelope)
Striped and Pelitic Group m^{2c} m^{2b} m^{2a}			Striped and Pelitic Group m^{2c} m^{2b} m^{2a}		Upper Pelitic Group mD	
Lower Psammitic Group m^1			Lower Psammitic Group m^1		Lower Psammitic Group mCb / METAMORPHIC BOUNDARY / mCa	
Unconformity			Thrust			
Outer Psammitic Group x''	SUB-MOINE Structural succession		Pelitic Group $m^{2a'}$	MOINE (core)	Lower Pelitic Group mB	MOINE (core)
Pelitic Group g'			Lower Psammitic Group $m^{1'}$		Central Psammitic Group (Base not seen) mA	
Intermediate Psammitic Group x'			Unconformity and Thrusts		Thrusts	
Hornblendic Group h / Subsidiary Pelitic Group g			Orthogneisses A	LEWISIAN	Orthogneisses and rare Paragneisses X	LEWISIAN
Central Psammitic Group x					Thrusts	

boundary " is used here to indicate the boundary mapped by Richey and Kennedy, with alterations as detailed later.

Inspection of Table 1 shows that the author departs to some extent from the previous views regarding the succession and relationship of the groups involved. The author's Upper Psammitic Group, Upper Pelitic Group, Pelitic Group of the Core, Central Psammitic Group, and Hornblendic Group (or Orthogneisses) do not differ from the same groups as defined in Richey and Kennedy (1939). The author's Lower Psammitic Group includes the Lower, Outer, and parts of the Intermediate Psammitic Groups of Richey and Kennedy and corresponds approximately to Kennedy's Lower Psammitic Group (Kennedy, 1955, Pl. XVII) but not in stratigraphic or structural position. It is generally divided into two parts by the core-envelope boundary. In this interpretation the Intermediate Psammitic Group of Richey and Kennedy is abandoned.

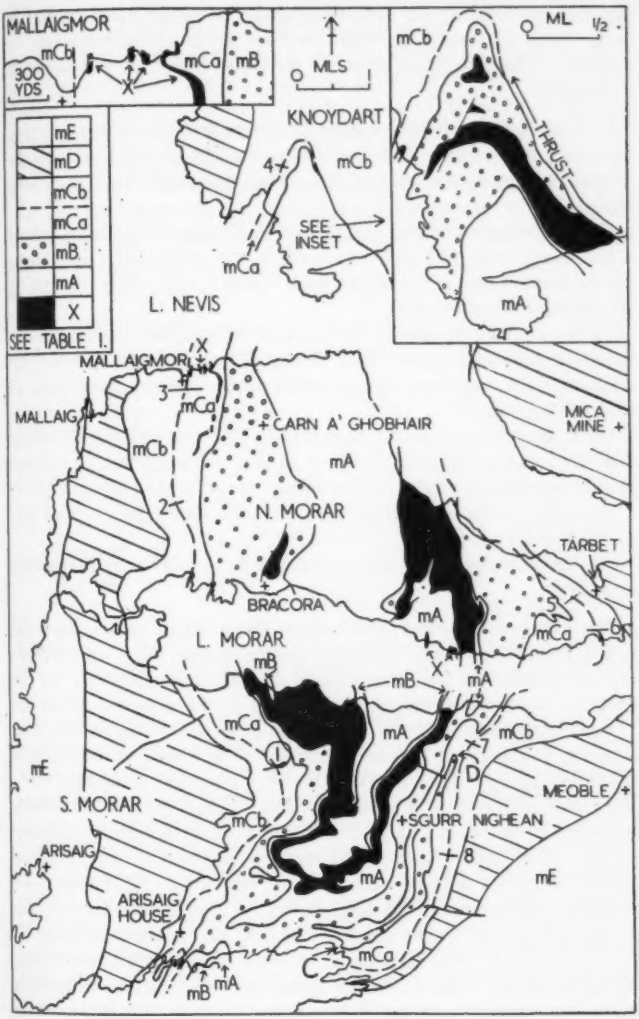
The writer wishes to thank Professor C. E. Tilley for suggesting the subject and for his continual help and criticism, Dr. J. E. Richey for his advice and help, Professor W. Q. Kennedy, Dr. A. G. MacGregor, and many others for their interest, help, and advice. He wishes to thank the Director of the Geological Survey for access to Morar and Knoydart rock slices included in the Survey's Scottish collection. The research was carried out during tenure of a D.S.I.R. research grant. Text-fig. 1 is published by permission of the Director of the Geological Survey; with the exception of the metamorphic boundary and minor amendments the geological lines are reproduced from six-inch sheets included in a simplified sketch-map, on a reduced scale, which forms Pl III. from *Bull. Geol. Surv. Gt. Brit.*, No. 2, 1939.

TEXT-FIG. 1.—The geology of the Morar district, Inverness-shire.

The Moine Series (see Table 1 for correlation):—

mE	Upper Psammitic Group	envelope
mD	Upper Pelitic Group	
mCb	Lower Psammitic Group	
mCa	Lower Psammitic Group	core
mB	Lower Pelitic Group	
mA	Central Psammitic Group	
X	Lewisian	

The entire margin of the Lewisian is thought to be a series of thrusts. Other thrust-boundaries exist in the east side of the core on Knoydart. The core-envelope boundary, generally lying within the Lower Psammitic Group, is marked by a dashed line. Traverses 1–8 across this boundary are marked by solid lines and labelled 1–8. Faults are also marked by solid lines; no dykes are mapped. B, C, and D correspond to the points labelled similarly on Pl. III, Richey and Kennedy, 1939. Point B is at Sgurr Nighean, S. Morar.



The Field Evidence and Structural Notes

The extent of the area under discussion is indicated on Text-fig. 1. The writer has intentionally departed from current conventions of

symbols representing the various subdivisions of the Moine Series in order to avoid confusion with the two previously existing sets of symbols. The map differs from that published by Richey and Kennedy as follows:—

(1) The Moine-Sub-Moine boundary is replaced by a metamorphic boundary (the core-envelope boundary) the line of which follows the original except on the west of the dome north and south of Arisaig House;

(2) The western boundary of the Pelitic Group of the core is moved about one-quarter of a mile eastwards at the point where it emerges on the north shore of L. Morar near Morar Lodge, thus eliminating an apparent unconformity;

(3) Some of the psammitic beds east of Sithean Mor, South Morar, have been allocated to the Pelitic Group of the core, which commonly contains thick psammitic bands where most typically developed, as near Mallaigmor;

(4) A redistribution of the Hornblende Gneisses and Pelitic Group has been made on Knoydart, where detailed mapping on aerial photographs was carried out; the changes do not affect the interpretation;

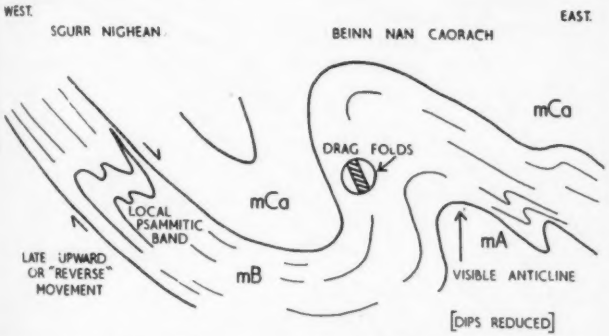
(5) Additional lenses of Hornblende Gneiss have been mapped east of Mallaigmor;

(6) The whole boundary of the Hornblende Gneiss Group is regarded as a thrust.

The first point above is discussed at more length below, while the fourth and fifth points do not require amplification. The second point is a consequence of the occurrence of new exposures made during the clearing of some densely wooded ground in the area east of Morar Lodge, while the third point follows on a re-examination of the area east and north-east of Sithean Mor, the evidence being summarized in the following. At the point D on Pl. III, Richey and Kennedy (1939), the structure is that of a simple *anticline*; hence the psammitic beds between D and B (loc. cit.) must belong to the Lower Psammitic Group (mCa). Anticlinal structure may also be observed at several points within the pelitic band between D and C (loc. cit.). At B (loc. cit.) a complex reversed drag fold develops within a psammitic band which passes laterally into pelitic schists. The drag fold is said to be reversed on account of the fact that the inner beds are dragged upwards, the whole occurring on the flank of an anticline. The reverse folds are only preserved where psammitic beds are interbanded with pelitic beds; elsewhere, north and south of B the pelitic beds have flowed and have a uniform outward dip. An interpretation of the structure is given on Text-fig. 2, showing the reverse fold within a normal drag fold. The sixth point will be discussed at greater length in a separate account of

the Lewisian; owing to lack of space no further evidence on this or the fourth and fifth points will be presented here.

In addition to the above, the author has noted a complete sedimentary-type gradation between the Pelitic Group of the core and the two Psammitic Groups on either side of it. This is best seen in N. Morar, but is consistent at all points in the area, though modified by tectonic attenuation in the area east of Sithean Mor. This evidence, taken with that given above and the recognition of the core-envelope boundary as a line of metamorphic origin, has led the author to believe that the Moine Series in Morar has a five-fold sedimentary sequence without either a visible base or top.



TEXT-FIG. 2.—A generalized west–east section through Sgurr Nighean, showing an interpretation of the core structure in the east part of S. Morar.

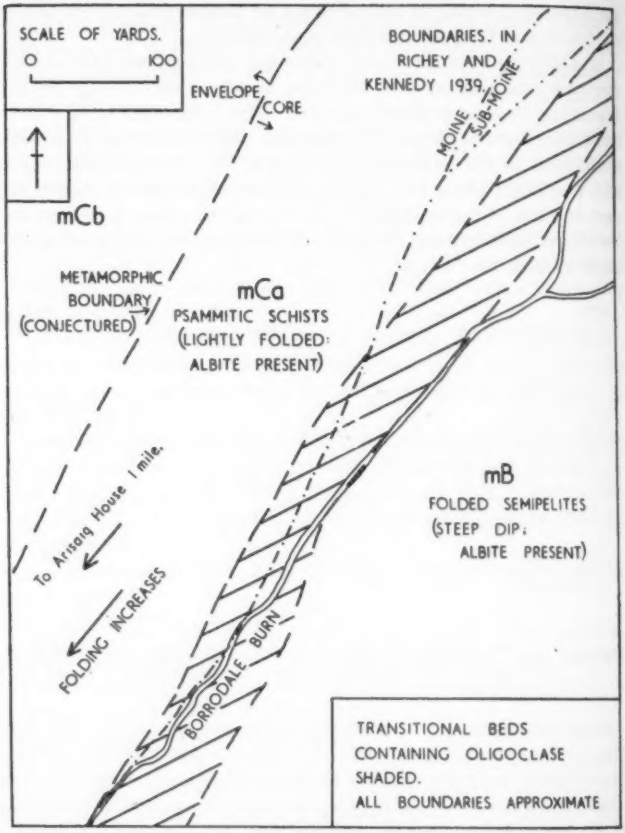
In earlier papers on the Morar structure general descriptions have been given of the rock types and of the nature and style of their folding. In particular, contrasts between the core and envelope have been discussed and accounts of the progressive metamorphism across this part of Scotland have appeared (MacGregor, 1952, p. 247; Kennedy, 1949). Information on the actual nature of the core-envelope boundary is, however, not available. In the field the boundary is of approximately uniform appearance except along the stretch within two miles of Arisaig House. It does not consist of a single mappable line, but exists as a zone within which a number of mappable features occur. Away from the Arisaig House area all changes are more or less gradational and are:—

(1) Degree of macro- and micro-contortion of foliation;

(2) The abundance and style of folding of quartz veins;

(3) The appearance of haematite flecks within the core.

The boundary as mapped marks the onset of a clearly visible oblique

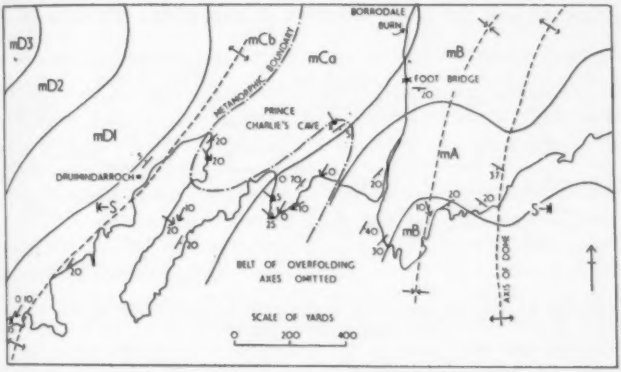


TEXT-FIG. 3.—A sketch-map of the geology of the upper reaches of Borrodale Burn.

foliation within the micaceous bands. Approximately coincident with this, macro-folding becomes ubiquitous and quartz veins (normally concordant to the foliation = bedding in the envelope) become discordant and folded. Folded quartz veins and haematite flecks usually appear a few hundred feet within the envelope as defined by oblique foliation. It must be emphasized that there are small areas within the envelope which show core characters, and vice versa; in addition, secondary folds, as north-west of Morar Lodge and north and west of Arisaig House are present, but in these areas the rocks do not show

core characters petrographically. One important but small area with core characters inside the envelope occurs 350 yards south-west of Morar Lodge, and may be seen on the road behind the boathouse.

In many places core and envelope rocks share similar styles of folding, notably near Mallaigmor and north-east and south-west of Arisaig House. To the north-east of Arisaig House the effect of the folding is to obscure the boundary in the field; on Pl. III in Richey and Kennedy (1939) the map shows an unconformity between Moine and Sub-Moine in this area, but the author prefers to move the boundary westwards on petrographical evidence (see Text-fig. 3). The position

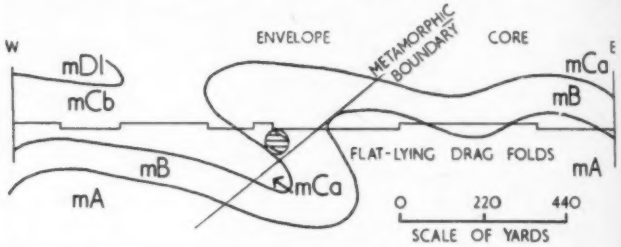


TEXT-FIG. 4.—The geology of the Arisaig House shore area. SS indicates line of section, Text-fig. 5. Arrows mark lineations (fold axes, pitch in degrees).

is complicated by the close approach of the boundary to the Pelitic Group, which is here mostly (near its west margin) garnetiferous and oligoclase-bearing, both envelope characters. These beds are slightly folded and are succeeded westwards, towards the envelope, by a band of albite-bearing epidotic psammitic beds which are relatively little folded. The latter character was responsible for the original mapping. Passing further west the secondary folding increases and obscures the fact that the mineralogy of the psammitic schists has become that of the envelope, oligoclase-bearing and with little or no epidote. The coincidence of a zone of little-folded rocks with the secondary fold zone north-west of Arisaig House made interpretation difficult in the absence of a detailed examination of the petrography. A map of this area is given as Text-fig. 3, on which all boundaries are conjectural, there being relatively few exposures. The position of the core-envelope boundary is not necessarily accurate, as it is based on very few

specimens, the entire evidence for its position being the microscopic mineralogy of the psammitic schists.

To the south-west of Arisaig House, in the vicinity of Prince Charlie's Cave, is another area where the core-envelope boundary is obscured in the field by the widespread development of secondary folds. The author's revised map of the area is given as Text-fig. 4. Exposures are not common away from the shore, so that the inland portions of the mapped boundaries are tentative. This area is mentioned by Richey and Kennedy (1939, p. 42) as one where " magnificent drag folds are developed in the attitude characteristic of inversion " and " the Lower Psammitic Group passes eastwards below the Sub-Moine core ". The author regards the latter statement as misleading; see Text-fig. 5.



TEXT-FIG. 5.—An east–west section along the Arisaig House shore. Vertical scale much exaggerated.

Exposures in the area of Text-fig. 4 are divisible into three groups :—

(1) To the south-west of the cave, with envelope characters in the field;

(2) To the south of the cave on the shore; pelitic and inverted beds much folded but not possessing the characteristic contortion of core rocks;

(3) To the south-east of the cave, folded and contorted pelitic and psammitic rocks with core characters.

In all areas, except at the west margin of the map and in a small area just west of the main axis of the dome, the dip of the bedding or foliation has an easterly component and is usually at low angles. This situation is unique on the west of the Morar area. There can be no doubt that the Lower Psammitic Group is disposed anticlinally near Druimindarroch and hence passes eastwards below the pelitic beds south of the cave. In order to avoid correlation of the latter with the Upper Pelitic Group (mD) the recumbent fold of Text-fig. 5 must be invoked, which corresponds closely with the observed " flat-lying drag folds ". Two sets of folds affect all rocks lying east of Druimindarroch

and west of the main dome axis, open folds on south-west plunging axes being the more prominent, while a " crinkling " or small-scale folding of the foliation plane plunges at low angles south-eastwards. The former decrease in amplitude westwards and barely affect the rocks at the west margin of the map, while the latter die out rapidly both east and west of the overfolded belt. Both sets of folds affect core and envelope equally. In addition, it may be noted that the Lower Psammitic Group to the west of the cave is more contorted macroscopically than the same group further to the south-west along the same line of strike. In the overfolded belt, specimens showing macrofolds at right-angles to microfolds (" crinklings ") may be obtained.

The Petrography and Mineralogy of the Envelope Schists

The envelope schists fall into two groups, those west and east of the core, each group containing psammitic and pelitic schists. It is necessary to note that the pelitic schists are not true " pelites " as they contain 2–3 per cent CaO by weight.

On the west of the dome a steady increase in metamorphic grade is observable on a traverse from the Rhu peninsula, west of Arisaig, eastwards towards the core near Arisaig House. The psammitic schists are quartz-microcline rocks generally with less than 20 per cent of plagioclase, epidote, micas, garnet, and calcite. At the extreme west coast the assemblage contains albite and common iron-rich epidote, which gives way eastwards to oligoclase up to An_{25} with rare iron-poor epidote, the three changes being gradational. Garnets are widespread, but not numerous; calcite rare. In the Upper Psammitic Group, thin pelitic strips contain the same assemblage with different mineral proportions; the co-existence of garnet and microcline (to be discussed elsewhere) being notable. The Pelitic Group of the envelope (mD) contains muscovite-biotite-garnet-quartz-oligoclase (near An_{25-30}) rocks with iron-poor epidote rare or absent. The Lower Psammitic Group has the same mineralogy as the easternmost beds of the Upper Psammitic Group, that is, quartz-microcline-muscovite-biotite-oligoclase-iron poor epidote-garnet (rare).

On the east of the dome the Lower Psammitic Group is much as on the west, while those parts of the Pelitic Group not involved in the " Injection Belt " are similar to that described above, though the plagioclase is more calcic, up to An_{32} occurring quite commonly. The metamorphic grade continues to rise eastwards.

The texture of those rocks in the above areas which contain sufficient quartz for the quartz grains to be in contact, has been described as " tesselate " (MacGregor, 1952, p. 245), that is, the quartz grains have smooth boundaries and are regularly shaped and of even size.

The Core-envelope Boundary

On approaching the core-envelope boundary, which generally lies in the Lower Psammitic Group, the pattern of a regular increase in grade eastwards is disrupted. Data concerning the mineralogy of specimens taken from a selection from a large number of traverses across the boundary are given on Table 2. The traverses are mapped on Text-fig. 1; the localities of the actual specimens do not always lie precisely on the lines mapped, but are not far from them. The specimens were randomly chosen as far as possible, but were invariably taken from normal psammitic beds. The general pattern is quite clear; traverse 4 on Knoydart illustrates the nature of the changes very well. The transition zone in the mineralogical changes generally lies to the envelope side of the mapped boundary, where it is recognizable, while some of the mineralogical changes invariably occur on the envelope side of the boundary, notably the appearance of haematite. The column labelled "haematite" refers to the occurrence or otherwise of small, translucent plates of haematite intergrown with muscovite. The quartz texture is self-explanatory; the iron content of epidote has been estimated by its birefringence; the composition of plagioclase was determined by means of the universal stage and Turner's curves (Turner, 1947).

The Tarbet traverses are exceptional in that the degree of variation of the plagioclase composition is small (but constant) and that garnet occurs only in the core specimens amongst those sampled. The garnet is, however, rare and being resorbed. One other mineralogical change occurs across the boundary which is not recorded on Table 2, that is, the type of alteration of garnet. Within the envelope, garnets are fairly free from alteration but where altered, are partly pseudomorphed by green or white chlorite. Within the core, with only two significant exceptions, the alteration is to biotite + muscovite + epidote without chlorite, actual pseudomorphs being common. The exceptions occur on the highest ground, one being described below.

The mineralogical changes towards the core across the boundary zone are summarized as follows:—

(*a*) the quartz texture gradually becomes sutured and inequigranular;
(*b*) plagioclase becomes more albitic, normal zoned plagioclase occurring in the envelope;
(*c*) epidote becomes more iron-rich;
(*d*) garnet does not usually appear in psammitic beds of the core and is altered to chlorite-free assemblages where it does; and
(*e*) haematite appears as intergrowths with muscovite.

Chemical change in the composition of plagioclase is very well illustrated on Table 3, which lists the plagioclase composition of the

TABLE 2.—PETROGRAPHIC DATA FOR SPECIMENS OBTAINED ON TRAVERSES ACROSS THE CORE-ENVELOPE BOUNDARY

Specimen number.	*Distance from boundary.*[1]	*Quartz texture.*	*Plag. An %.*[4]	*Epidote iron content.*[5]	*Haematite.*	*Garnet.*	*Remarks.*
S. of L. na Ba Glaise, traverse 1							
628	30 yds. W.	tesselate	16	low	rare	nil	envelope
629	15 yds. W.	semitess	20	medium	rare	nil	
630	5 yds. W.	semitess	20	medium	present	nil	transition zone
631	20 yds. E.e	sutured	10	high	present	nil	
N. of L. a Bhada Dharaich, traverse 2							
485	nil	semitess	14	medium	present	nil	
486	10 yd. E.	semitess	2	high	present	nil	core
Mallaigmor, traverse 3							
325	300 yd. W.	tesselate	10	medium	nil	rare	
326	190 yd. W.	tesselate	<20	medium	present	rare	envelope
327	30 yd. W.	tesselate	17–24 [3]	medium	nil	rare	
328	70 yd. E.	sutured	9	high	rare	nil	
329	160 yd. E.	sutured	2–0 [3]	high	rare	nil	
330	300 yd. E.	sutured	—	high	rare	nil	core
Airor track, traverse 4							
536	120 yd. W.	tesselate	28–10	medium	present	rare	envelope
537	30 yd. W.	tesselate	17–6	low	nil	rare	
541	20 yd. W.	tesselate	13	medium	rare	rare	transition zone
542	60 yd. W.	tesselate	13	medium	nil	nil	
538	10 yd. E.	semitess	5	medium	rare	nil	
539	80 yd. E.	semitess	13	medium	nil	nil	
540	30 yd. E.	sutured	1	high	present	nil	core
Tarbet, traverses 5 and 6							
568	100 yd. E.	tesselate	24–12	medium	present	nil	
571	160 yd. E.	semitess	20	medium	nil	nil	
576	140 yd. E.	semitess	19	medium	nil	nil	
577	190 yd. E.	tesselate	26	low	nil	nil	envelope
578	330 yd. E.	tesselate	25	nil	nil	nil	
569	120 yd. W.	semitess	10	high	present	rare	
570	10 yd. W.	semitess	17	medium	rare	nil	
572	150 yd. W.	sutured	20	medium	rare	nil	
574	210 yd. W.	semitess	19	high	rare	rare	core
575	50 yd. W.	semitess	19	medium	nil	nil	
L. a Ghriasaiche, traverse 7							
615	30 yd. E.	tesselate	30–18	medium	present	nil	
616	10 yd. W.	tesselate	26–23	high	present	nil	envelope
617	18 yd. W.	tesselate	8	medium	rare	nil	
618	24 yd. W.	semitess	4	medium	nil	nil	
619	50 yd. W.	semitess	20–11	high	nil	nil	core
Bealach a Mama, traverse 8							
659	nil	semitess	29	low	nil	nil	envelope
658	30 yd. W.	semitess	17–10	high	present	nil	
657	31 yd. W.	semitess	18–6	high	present	nil	core

[1] Distance from boundary as mapped by degree of contortion. [3] 17–24 indicates reverse zoning in plagioclase. [3] 2–0 indicates normal zoning in plagioclase.
[4] All plagioclases composition ± 3 per cent An. [5] Epidote iron content: low, 0–7 wt. per cent Fe_2O_3; medium, 7–10 per cent; high, > 10 per cent.

specimens nearest to the mapped boundary on either side at a large number of points, listed clockwise round the boundary starting on the west of the core at the watershed in S. Morar. The four traverses on the east of the dome (Tarbet-Arnipol inclusive) have higher anorthite contents in the envelope schists than the remainder, in keeping with the general easterly increase in metamorphic grade. The mean value of the compositions given is An_{13} and this value has been taken by the author as the best criterion for the position of the boundary in the parts of S. Morar where the boundary is obscured in the field by secondary folds.

The Arisaig House District

The available data concerning the mineralogy of the folded schists from south of Arisaig House (see above, Text-figs. 4 and 5 and text)

Table 3.—Anorthite Content of Plagioclase in Specimens obtained nearest to the Mapped Boundary

Place.	*Nearest mCa (core).*	*Nearest mCb (envelope).*	*Yards apart.*
S. of L. na Ba Glaise	10	20	25
N. of L. na Ba Glaise	12	21	30
S. of L. a Bhada Dharaich	9	15	100
N. of L. a Bhada Dharaich	2	14	10
Allt a Bhainne	8	4	60
Mallaigmor	9	17	100
Sandaig	nil	12	60
Airor Track	5	13	70
N. of Lagan Bridge	6	11	50
Tarbet (averages)	17	23	—
L. a Ghriasaiche	8	23	8
Bealach a' Mama	10	29	30
Arnipol	7	29	30
W. of Borrodale Viaduct	4	15	90
N. of Borrodale Viaduct	11	22	70

are summarized on Table 4. The specimens are listed in groups corresponding with the divisions established in the field. In the first group one, 843, possesses core characters, while the remainder are of transitional type (i.e. in the envelope as usually mapped in the field). 843 comes from the shore 400 yards west of Prince Charlie's Cave (see Text-fig. 4) and lies at the west side of an area extending north-east past the cave containing isolated exposures of folded rocks of core type. 844 comes from the cave itself.

The second group, from the shore south of the cave, has characters conforming to envelope or transitional beds (specimens S 32781, etc., are from the Scottish Survey Collection) but occurs within a highly folded zone.

The third group, to the south-east of the cave, all possess core characters (except 846, of transitional type) and include specimens

taken from the Central Psammitic Group and the outcrop of the core Pelitic Group on the shore of Borrodale Burn. It is concluded that the metamorphic boundary, which cannot be mapped in the field because of the secondary folding, actually transgresses the geological boundaries as indicated on Text-fig. 4 and shown diagrammatically on Text-fig. 5. It may be noted that the Pelitic Group of the core actually

TABLE 4.—PETROGRAPHIC DATA FOR SPECIMENS OBTAINED NEAR ARISAIG HOUSE (MAP, TEXT-FIG. 4)

	Quartz.	*Plagioclase Anorthite percentage.*	*Epidote iron-content.*	*Haematite.*	*Garnet.*
(i)					
860	tess	14 ± 4	medium	rare	nil
859	tess	18 ± 3	medium	rare	nil
858	tess	14 ± 3	medium	nil	rare
844	semitess	15 ± 2	medium	nil	nil
843	semitess	6 ± 4	high	nil	nil
(ii)					
857	tess	18 ± 3	low	nil	nil
S 32781	n.d.	n.d.	low	nil	nil
S 32783	tess	n.d.	low	nil	present
856	tess	16 ± 3	medium	nil	nil
855	semitess	16 ± 4	medium	nil	nil
(iii)					
845	sutured	<10	high	present	nil
846	semitess	15 ± 3	medium	rare	rare
853	sutured	< 8	high	present	rare
854	sutured	< 6	high	present	nil
849	sutured	< 7	high	present	nil
850	sutured	<10	high	present	nil
S 32763	sutured	n.d.	high	present	nil
S 36792	sutured	n.d.	high	present	nil
847	sutured	nil	medium	present	nil
848	granulated	<11	high	present	nil
S 36304	granulated	n.d.	medium	nil	nil

crosses into the envelope in this interpretation. It may be concluded that the folding did not occur after the metamorphism.

THE CORE SCHISTS

All psammitic schists of the core have the assemblage quartz-microcline—albite (less than An_{15})—iron-rich epidote—muscovite with haematite-biotite; accessories, sphene and haematite.

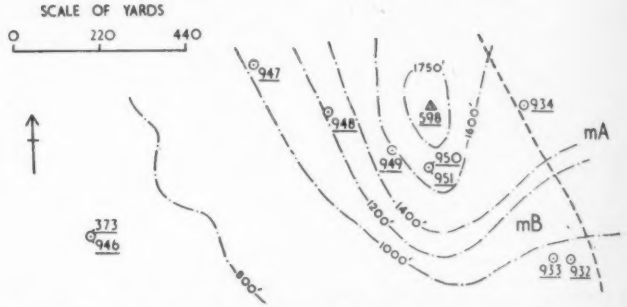
The pelitic schists, which are quite different lithologically from the envelope Pelitic Group and do not contain calc-silicate bands or heavy mineral segregations, may be divided into three groups:

(i) muscovite—biotite—oligoclase (*c.* An_{30})—garnet—quartz;

(ii) muscovite with haematite—biotite—oligoclase (An_{15-20})—low-iron epidote—garnet—quartz;

(iii) muscovite with haematite—biotite—albite (An_{0-15})—high-iron epidote—quartz ± microcline.

The three groups grade into one another. Group (i) is very restricted and is found only near the core-envelope boundary (see Text-fig. 3) or on the highest ground. By increase in epidote at the expense of anorthite and by the partial replacement of garnet by muscovite and biotite this group grades into (ii). Garnet is antipathetic to microcline (cf. the Upper Psammitic Group of the west coast described above) and to muscovite with haematite. Group (ii) grades into group (iii) without microcline by a continuation of the groups (i) to (ii) changes but the status of group (iii) with microcline may not be of the same character. The destruction of once-idiomorphic garnet coupled with the other changes observed across the core-envelope boundary suggest retrogressive metamorphism within the core accompanying the deformation which produced the folding responsible for the mapped core-envelope boundary. The mechanism of production of a relatively sharp boundary will be discussed in a later communication.



TEXT-FIG. 6.—Localities and heights of specimens from near Carn a' Ghobhair, N. Morar.

THE CARN A' GHOBHAIR AREA

The mountain Carn a' Ghobhair in N. Morar is made of the core Pelitic Group, the mineralogy of which, on the higher reaches of the mountain, is rather exceptional. Text-fig. 6 shows the localities of specimens described in Table 5. The four specimens in group (*a*) all come from the ground below 1,000 feet and belong to group (iii) of the types listed above from the Pelitic Group. The specimens in group (*b*), Table 5, are listed in order of ascending height, and form an ascending sequence from group (iii) through (ii) to (i). The summit specimen (598) is a rock of envelope type mineralogically, *including the alteration of garnet to green chlorite*. In 949 the garnet is being altered to epidote + muscovite + biotite; in 950 and 951 the

garnets, rare and resorbed, are altered to the same assemblage as 949, except that the biotite, instead of being the usual homogeneous brown colour (as it is throughout the core), is brown margined by green, the green part possessing biotitic and not chloritic optical properties. The green biotite appears to represent a transitional stage between the brown biotite and chlorite stages of alteration of garnet. As schists of group (i) are otherwise confined to the margin of the core and the alteration of garnet to chlorite is an envelope feature, it is thought that

TABLE 5.—PETROGRAPHIC DATA FOR SPECIMENS OBTAINED NEAR CARN A' GHOBHAIR (MAP, TEXT-FIG. 6)

	Plagioclase Anorthite percentage.	*Epidote iron content.*	*Garnet.*	*Haematite.*
(*a*)				
373	15 – 1	medium	nil	rare
946	10 ± 3	medium	nil	present
932	<10	high	nil	present
933	13 ± 2	medium	nil	present
(*b*)				
947	<11	medium	nil	present
948	14 ± 2	high	nil	present
949	20 ± 3	low	present	nil
950	11 – 18	low	common	nil
951	22 – 9	low	common	nil
598	23 ± 2	low (rare)	common	nil

the summit rocks of Carn a' Ghobhair are approaching the domed metamorphic boundary. If they are in fact doing so, the boundary is less domed than the rocks, agreeing with the evidence of transgression found at Arisaig House. On the highest ground in S. Morar, analogous events within the Lewisian accord with this interpretation.

CONCLUSION

The Moine schists of Morar are thought to have been affected by a progressive and a subsequent retrogressive metamorphism on the basis of general petrographic evidence. Only the core schists were affected on a major scale by the retrogressive metamorphism, the outer limit of which is marked by the core-envelope boundary which, though generally mappable in the field as a line, is actually a zone of transition of variable width as now exposed. This zone transgresses the geological boundaries. It is to be noted that the metamorphic state of the core schists corresponds mineralogically with that of the low-grade envelope schists on the extreme west of Morar. In this account the petrographic details for the above conclusions have been given; it forms an introduction to the discussion of the chemistry of the minerals of the schists and to a discussion of the Lewisian which will be published later.

REFERENCES

BAILEY, E. B., 1939. Caledonian tectonics and metamorphism in Skye. *Bull. Geol. Surv. Gt. Brit.*, No. 2, 46–62.

CLOUGH, C. T., 1907. *In* The geological structure of the north-west Highlands of Scotland. *Mem. Geol. Surv. U.K.*

—— 1910. *In* The geology of Glenelg, Lochalsh and the south-east part of Skye. *Mem. Geol. Surv. U.K.*

KENNEDY, W. Q., 1949. Zones of progressive regional metamorphism in the Moine Schists of the Western Highlands of Scotland. *Geol. Mag.*, lxxxvi, 43–56.

—— 1955. The tectonics of the Morar anticline and the problem of the north-west Caledonian front. *Quart. Journ. Geol. Soc.*, cx, 357–390.

MACGREGOR, A. G., 1948. Resemblances between Moine and Sub-Moine metamorphic sediments in the Western Highlands of Scotland. *Geol. Mag.*, lxxxv, 265–275.

—— 1952. Metamorphism in the Moine nappe of Northern Scotland. *Trans. Edin. Geol. Soc.*, xv, 241–257.

RICHEY, J. E., 1948. Pre-metamorphism cleavage in the Moine schists of Morar, Inverness-shire. *Trans. Edin. Geol. Soc.*, xiv, 200–219.

—— and W. Q., KENNEDY, 1939. The Moine and Sub-Moine series of Morar, Inverness-shire. *Bull. Geol. Surv. Gt. Brit.*, No. 2, 26–45.

TURNER, F. J., 1947. Determination of plagioclase with the four-axis universal stage. *Amer. Min.*, xxxii, 389–410.

DEPARTMENT OF GEOLOGY AND MINERALOGY,
UNIVERSITY MUSEUM,
OXFORD.

The Quihita-Cunene Layered Gabbroic Intrusion of South-west Angola

By P. Stone and G. M. Brown

(PLATE VI)

Abstract

The detailed examination of rocks recently collected from a traverse and from isolated points within this intrusion has led to the conclusion that it is a large, layered intrusion showing rhythmic and cryptic variation, and comparable in many respects with the Great Dyke of Southern Rhodesia.

I. Introduction

THE area shown on the map (Text-fig. 1) was, geologically, the least known and worse-mapped part of Angola before the Companhia de Diamentes de Angola started prospecting in 1930, previous expeditions in the region being of a purely exploratory nature. In 1933, however, Mouta and O'Donnell published a geological map of Angola, which included a large area of gabbro and anorthosite, cutting the Huronian granite and other rocks of the basement complex in the region of the high scarp and interior plateau. In the same year Beetz published an account of the geology of part of South-west Angola, including a brief description of the gabbro-anorthosite complex, reference being made to a few earlier papers in which the presence of basic rocks in this region was recognized.

The Oxford University (Undergraduate) Expedition to Angola (1954) included one geologist (P.S.), who was able to make an intensive collection across the probable axis of the gabbro intrusion, in an east–west direction from Pocolo to Gambos (Text-fig. 2), and a less complete collection to the south near the Rio Cunene. The road from Pocolo to the railhead at Chiangue, and thence to Gambos, was mapped by plane-table, and most of the specimens were collected along this road, across the Rio Caculovar to Gambos and the prominent landmark, Tonga Tonga. Further specimens to the north-west and south-west of Gambos were collected along the Chibia–Humbe road. The Expedition then approached the Rio Cunene along a track following the Rio Mupelale, but detailed collection was inhibited in this region owing to transport difficulties. The present paper is the outcome of a petrographic study of this collection, chiefly represented by the Pocolo-Gambos traverse.

II. The Results of Earlier Investigators

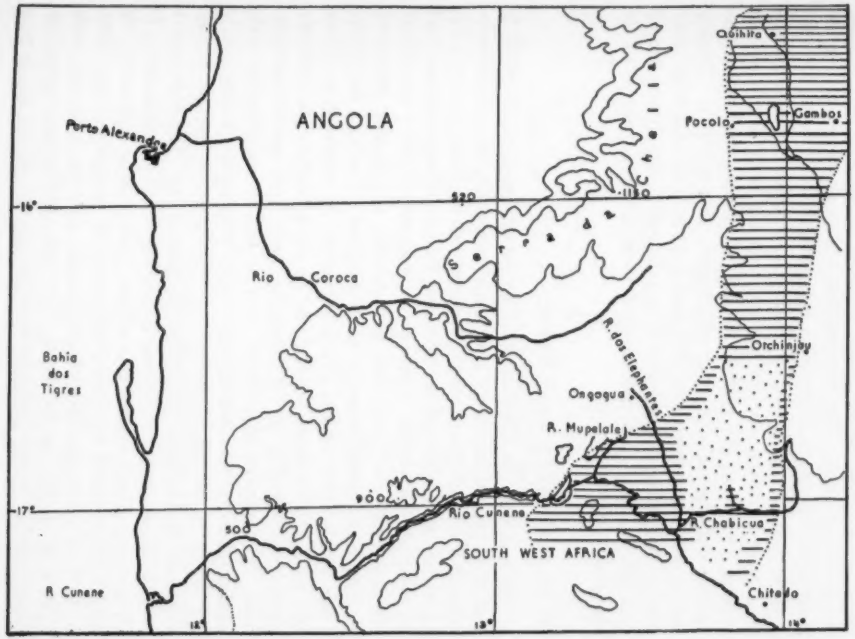
As much of the previous work is difficult of access to English

readers, references to the basic intrusion being scattered widely amongst papers of more general interest, a brief review of the relevant points is included at this stage.

The gabbro-anorthosite mass of South-west Angola is a very large body elongated north–south and measuring 300 kilometres by at least 25 kilometres. The investigations which preceded that of Beetz (1933) were usually confined to small parts of the intrusion, related to the general geology of the district. Souza (1916), describing specimens collected by others, recognized a great extent of gabbros, troctolites, norites, anorthosites, and diabase, intruding the granites and cut in turn by alkali-syenite and pegmatites. Vageler (1920) mentioned that the ranges bordering the Kalahari are mostly of anorthosite-gabbros, becoming fine-grained and less feldspathic towards their contact. He recognized a north–south orientation of the gabbro mass and commented upon large lenses of ilmenite-magnetite towards the centre of the intrusion. (Vageler was of the opinion that the gabbros represented the plutonic equivalents of Karroo lavas). Faber (1926) investigated the intrusion between Chibia (65 km. north-west of Quihita, Text-fig. 1) and Gambos, describing briefly olivine-gabbro and anorthosite, but also a granophyre containing quartz with undulose extinction (cf. Pl. VI, fig. 3).

Beetz (1933) studied the intrusion in the southern part, where it crosses the Rio Cunene. There he recorded frequent mineral banding, the mass being predominantly anorthosite with gabbro forming narrow bands on both sides. The petrographic description is very brief, the anorthosites being treated separately from the gabbros. The former are described as coarse, hypidiomorphic, and consisting of labradorite and, less commonly, bytownite. The gabbros are described as coarse; the feldspar is labradorite, bytownite, and anorthite, the orthopyroxene is enstatite, bronzite, and hypersthene, and monoclinic pyroxene is chiefly diallage (this variable mineralogy, if true, indicates appreciable variation within the " gabbros "). The slight crushing recorded in the anorthosites is no more extensive than that recognized in the gabbros. Inclusions of country rock, classed as " mangerites " (cf. Pl. VI, fig. 3), are briefly described and identified with the alkali-syenite " intrusions " of Souza. Beetz concludes that the mass is Pre-Cambrian in age and comparable with the large anorthosite masses of North America and Scandinavia.

Recently Assunçao, Mouta, and Brak-Lamy (1952) have given brief descriptions of the basic rocks, and divided them into a gabbro and an anorthosite series, sub-dividing the latter into plagioclasites, norites, troctolites, and hyperites. The gabbros are sub-divided by the presence or absence of olivine, but the olivine-free gabbros are said to be marginal to the more calcic olivine-gabbros.



TEXT-FIG. 1.—Map of the Quihita-Cunene gabbroic intrusion (horizontal lines). Reproduced from Beetz (1933, pl. vi) with modifications. Scale 1:1,000,000. Contours at 500 m. and 1,000 m. Dotted areas: Cover of red desert sandstone. Unshaded areas: Basement complex, chiefly granite-gneiss. Enclosed unshaded areas: foundered blocks of ? roof, and contaminated zones.

III. The Pocolo-Gambos Traverse

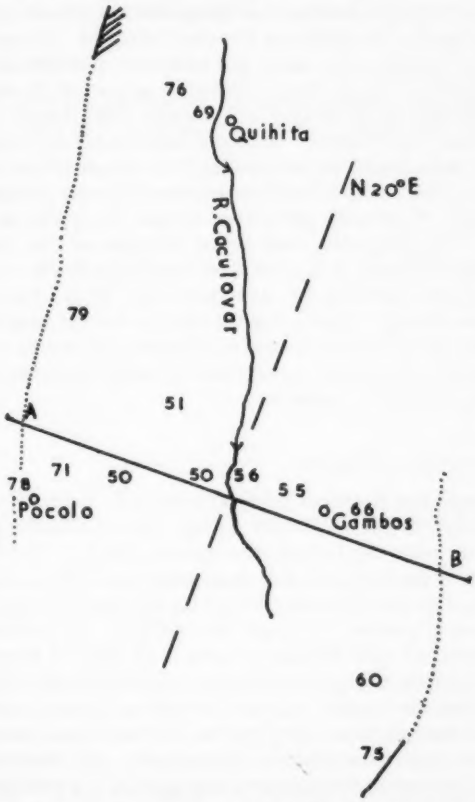
(*a*) *Introduction*

The work of previous investigators has established that the intrusion consists of a varied assemblage of basic rocks, confined to an elongate area measuring 300 km. (n–s) × 25 km., intruded largely into Pre-Cambrian metasediments and granites. No attempt has been made, however, to explain the variation in the basic rocks, or to examine the nature of the variation. The aim of the present investigation was to examine the structural, textural, and mineralogical relationships of the basic rocks within the intrusion, in order to better understand their mode of formation. For this purpose a traverse was made from the western to the eastern contacts, in the northern sector where the width is close to 40 km. The eastern contact was reached by working south-west from Gambos. In the western part the relief is low and the weathering severe, the gabbro generally appearing as isolated hummocks, but exposures improve with higher relief eastward. Contacts with country rock were never obtained, but could be established within one or two miles of exposures of granite-gneiss, the typical country rock (Pl. VI, fig. 3), and the marginal facies (usually melanocratic) of the gabbro intrusion.

(*b*) *The Basic and Ultrabasic Rocks*

The rocks of the intrusion have an overall gabbroic character, in that they are coarse-grained and consist essentially of a calcic plagioclase feldspar and clinopyroxene, together with accessory olivine, orthopyroxene, and an iron-oxide. However, the assemblage is characterized by rhythmic and cryptic layering, two properties which make difficult the classification of every rock as a gabbro, and equally difficult the allocation of separate names to each specimen. Although the layering is not visible in the field as a major structure, for exposures are not on this scale, the rhythmic layering can be recognized by the variability in the proportions of the constituent minerals from specimen to specimen. The absence of evidence such as gravity stratification makes distinction between rhythmic and cyclic layering impossible, but as the former is the more common type of repetitive layering in basic intrusions that name is adopted in this description. Mineral variation is such that marginal rocks may contain up to 70 per cent olivine, or 70 per cent clinopyroxene plus olivine, poikilitically enclosed by the mineral(s) absent as a primary precipitate (cf. Wager and Deer, 1939, p. 127), in the former case plagioclase and clinopyroxene, and in the latter case plagioclase (Pl. VI, fig. 1). Inwards from the margins the proportion of olivine decreases, but although it may be absent from

some specimens it is present in adjacent specimens, indicating the type of rhythm frequently attributable to crystal accumulation under the effects of gravity. Similarly the pyroxene proportions may vary in



TEXT-FIG. 2.—The Pocolo-Gambos traverse (A–B), approximately perpendicular to the intrusion axis, indicating the position of most of the rocks from which feldspar compositions were determined. Figures indicate percentage An in plagioclase. Boundary (dotted) inferred in the absence of exposed contacts. Scale 1:500,000.

relation to plagioclase, resulting in melanocratic and leucocratic rocks from adjacent exposures. The rocks rich in feldspar may contain poikilitic pyroxene, poikilitic olivine (Pl. VI, fig. 2) or both. The presence of pure feldspar layers suggest the removal of interstitial liquid, through

a diffusion process similar to that outlined for the pure feldspar layers in Rhum (Brown, 1956, p. 37). The wide spread of the feldspathic rocks in the centre of the intrusion may be due either to the low dips near the centre (the synclinal shape exaggerating the thickness even further), or to the predominance of feldspathic layers, such as found in the upper parts of the Stillwater Complex, Montana. In any case the presence of melanocratic bands and occasional poikilitic olivines in the feldspathic rocks is strong evidence in support of the suggestion now made that we are dealing with a single differentiated intrusion, and that the " anorthosites " and more melanocratic (on the average) marginal rocks should not be considered as separate intrusions.

In view of the variety of rock-type produced through variation in the proportions of primary precipitate olivine, clinopyroxene, orthopyroxene, and plagioclase, and in the character of the crystallized interprecipitate liquid, it has not been thought profitable to describe each rock-type separately (cf. Assunçao *et al.*, 1952). The evidence given above strongly suggests that the mass is a layered basic intrusion (cf. Wager, 1953), and that a trend of differentiation may be traceable, as with larger intrusions, in the proportions, compositions, and textures of the major constituent minerals.

(*c*) *The Plagioclase Feldspars*

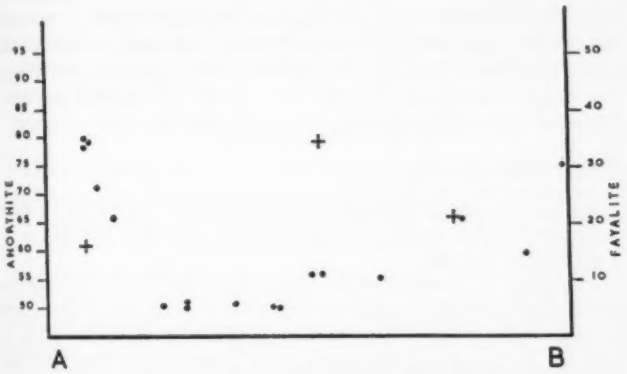
The plagioclase is present either as subhedral, discrete crystals, in the rocks rich in this mineral (Pl. VI, fig. 2) or as anhedral, poikilitic plates in the rocks rich in the ferro-magnesian minerals (Pl. VI, fig. 1). Such a habit has been described from other basic intrusions (Wager and Deer, 1939, and Brown, 1956), and can best be explained by reference to a process of crystal accumulation, the textures being chiefly dependent upon whether the mineral (in this case feldspar) is a primary precipitate enlarged by interprecipitate growth or produced entirely from the crystallization of the interprecipitate liquid. The amount of feldspar varies from 5 per cent in the melanocratic rocks to 100 per cent in the monomineralic feldspar rocks. The discrete crystals are unusually large in this intrusion, ranging from a few millimetres to several inches in length. A few rare crystals attain a length of two feet. The plagioclase is usually fresh, twinned on both albite and pericline laws, and commonly shows normal zoning.

The composition, determined by measurement of maximum extinction angle, $X'\wedge 010$, using the Universal Stage, ranges from An_{75-80} in rocks from each margin to An_{50} in rocks near the centre. The change, despite the limited number of exposures, is gradual, as can be seen in Text-fig. 3. Such a change is indicative of fractional crystallization of a basic magma by bottom-accumulation. In many of the more

calcic examples the feldspar crystals are strongly zoned (normal discontinuous), in which case the composition of the core is used for comparison with the larger primary precipitate crystals of other rocks, the rim being merely an indication of the changing composition, on cooling, of the interprecipitate liquid.

(*d*) *The Olivines*

Olivine is present in most of the rocks collected, averaging 20 per cent by volume (but reaching 70 per cent) in those from nearer the



TEXT-FIG. 3.—Plotted compositions of plagioclases (circles = An percentage) and olivines (crosses = Fa percentage) in Pocolo-Gambos traverse (A–B shown in fig. 2).

western and eastern margins, but down to 5 per cent or a few scattered grains in the central part of the intrusion. The crystals are euhedral and discrete where abundant (Pl. VI, fig. 1), but subhedral and irregular where present in small amounts (Pl. VI, fig. 2). The olivines are very fresh, serpentinization being usually confined to a few narrow cracks in each grain.

The compositions were determined, where possible, by measurement of $2V\alpha$. The marginal olivines have the lower fayalite content, such as Fa_{10} associated with plagioclase An_{76}, and Fa_{20} associated with plagioclase An_{66}. Towards the centre of the intrusion the olivine fayalite content increases, Fa_{35} being associated with plagioclase An_{55} (see Text-fig. 3). There is thus a progressive iron enrichment in the olivines in the same direction as the sodium enrichment in the plagioclase feldspar.

(e) The Pyroxenes

The orthopyroxene is rare in the marginal, olivine-rich rocks, usually occurring as narrow rims fringing the olivines. Towards the centre the average content of orthopyroxene increases, and subhedral crystals may comprise up to 20 per cent of the rock. Several of the latter have a composition, estimated from $2V_\gamma$, of about $Mg_{55}Fe_{45}$, and all contain exsolution lamellae of augite || (100). In view of their composition, it is of interest that none of the orthopyroxenes show evidence of inversion from pigeonite.

The clinopyroxene may be present as discrete, primary precipitate crystals forming up to 70 per cent of the rock, near the margins, or as large poikilitic plates where less abundant and interprecipitate, towards the centre. Compositions have not yet been measured, in view of the difficulty of correlating optical properties and composition, but $2V_\gamma$ of the marginal specimens averages 50°. Exsolution lamellae are rare, but a few fine lamellae of orthopyroxene || (100) have been detected.

(f) The Other Minerals

Apatite becomes increasingly common towards the centre, where it occurs as euhedral crystals, average diameter 0·3 mm., forming up to 1 per cent of the rock.

Iron oxides are absent from the rocks nearer the margins, but become increasingly common towards the centre of the intrusion, particularly in those rocks in which apatite is prominent. In addition, lenses of iron ore are found near the centre, some prospecting having disclosed that these bodies can be quite large.

Hornblende, as large, green, poikilitic plates is confined to a few of the marginal differentiates, where contamination with country rock is suspected. A few small grains, apparently of secondary origin, are found in some of the more central rocks. Biotite is not uncommon, and occurs chiefly as rims to the iron oxides in some of the inner rocks.

IV. The Central Roof

South of Chiangue (between Pocolo and Gambos, see Text-fig. 1) is a small hill formed by an outcrop of rocks, about one mile square, foreign to the basic rocks of the intrusion on each side of it. These foreign rocks have a gneisso-granitic texture and the lighter variety consists of about 50 per cent quartz, 40 per cent feldspar, and up to 10 per cent orthopyroxene, iron oxide, and apatite. The quartz invariably shows strained shadowy extinction and frequently serrated crystal boundaries (Pl. VI, fig. 3). The feldspar consists of turbid microcline and clearer oligoclase with well-developed myrmekite texture. The darker varieties are rich in aggregates of orthopyroxene, green

amphibole, and biotite, an assemblage sometimes contributing to xenoliths in the gabbroic rocks of the intrusion.

Since the prevailing country rock surrounding the Quihita-Cunene gabbro in this region is granitic-gneiss, it is suggested that this patch of foreign rock represents part of the original roof of the intrusion, though in view of the absence of associated chilled gabbro and late differentiates it must have foundered into its present position. Such rocks were often classed as intrusive syenites and granites by earlier investigators, an error which led to variety in the interpretation of the shape of the intrusion. For example, the syenite near Pocolo mentioned by earlier writers was not located by the writers, and it may well have been the block of roof mentioned above. As a result the boundary of the Quihita-Cunene basic intrusion has now been placed to the west of Pocolo and not to the east (Text-fig. 1), a lateral displacement of about 12 km.

V. The Southern Region

Although Beetz (1933, pl. vi) showed the basic intrusion continuing southward from Pocolo–Gambos for about 80 km., to Otchinjou, and outcropping along the Rio Cunene a further 75 km. south-west of Otchinjou, at least 50 km. of the latter-mentioned 75 km. is shown on the map to be occupied by a large outcrop of earlier (? Huronian) Chella alkali-granite. This earlier granite mass seems to divide the gabbroic body into separate and distinct northern and southern intrusions and this area was therefore visited in order to see if such was the case. In fact, east of the Rio dos Elephantes the gabbro is still found, extending east as far as the Otchinjou–Chitado road (Text-fig. 1), and everywhere capped by up to 700 ft. of almost horizontal red feldspathic sandstones which possess many characters suggestive of continental deposition. It is possible that Beetz misidentified these sandstones from a distance, or through an error in examining earlier collections, for there is no evidence in his accounts of having visited the area. In view of this fresh evidence it is clear that the intrusion extends unbroken for over 200 km. from Capelongo, through Quihita, to the Rio Cunene. It may be that the reason for the apparently random weathering of the gabbroic rocks further north, where small hills of fresh rock lie close to deeply weathered areas, is that the kopjes were protected by a capping of sandstone (or original roof) for longer than the other deeply weathered parts.

Large areas of pink and white rock in the southern region, particularly above the Montenegro Falls on the Rio Cunene, appear to be gabbroic rocks greatly altered by a late-stage deuteric process. This alteration is localized, and certainly not a surface alteration, for it is not uncommon to see the white rock in the deeper gullies overlain by

the dark unaltered gabbro. This phenomenon is particularly well seen in the Zebra Mountains in South-west Africa, about 10 km. south of where the Rio Mupelale joins the Rio Cunene, the white rock extending, in the gullies, halfway up the mountainsides.

In this southern region the western edge of the intrusion appears to be in the vicinity of the Rio Mupelale, where a few specimens were collected very similar to the ultrabasic differentiates found near the margin at Chibia (65 km. north-west of Quihita, Text-fig. 1), in the northern region. The olivine in one of these rocks is between Fa_{12} and Fa_{16}, while the plagioclase in a rock further from the margin is An_{68}. Gabbro collected from nearer the central axis of the intrusion, 25 km. east of the Rio dos Elephantes, contains a plagioclase with An_{58}, and is texturally very similar to central specimens from the northern traverse. Again, as in the north, olivine becomes less frequent but does not disappear, inwards towards the central axis.

VI. Conclusions

The work of earlier investigators in the region between Chibia and the Rio Cunene, South-west Angola, disclosed the presence of a great variety of basic igneous rocks confined to an elongate zone approximately 300 km. (north–south) by 25 km. (east–west). The maps indicated the possibility of separate intrusions, as did the classification of the rocks into distinct groups; in particular the gabbroic group was divided from the anorthosite and Beetz (1933) drew a comparison between the anorthosites of Angola and the massive anorthosites (unlayered and relatively uniform in composition) of Scandinavia and North America.

The present investigation has shown that the Quihita-Cunene gabbroic mass is probably a single intrusion of basic magma which has differentiated to produce a series of layered rocks of contrasted compositions. Rhythmic layering is detectable in the modal and textural relationships of adjacent rocks, while cryptic layering is shown chiefly by the progressive change in composition of the olivines from Fa_{16} to Fa_{25} and of the plagioclases from An_{80} to An_{50}, inwards from the margins of the intrusion. Analogies can be drawn with several other layered basic intrusions with reference to the variation in mineral compositions, mineral associations, and rock textures, but the analogy which immediately springs to mind is with the Great Dyke of Southern Rhodesia (Hess, 1950). Both intrusions are long, almost parallel, dyke-like masses striking NNE–SSW and intruded into Pre-Cambrian granite and gneiss. Each consists of narrow marginal zones of more melanocratic ultrabasic rocks passing gradually inwards into a broader central zone of more leucocratic basic rocks. The rocks of both the

Quihita-Cunene and the Great Dyke show strong evidence of cryptic and rhythmic variation, usually attributed to crystal accumulation under gravity, although in the former case olivine is the chief melanocratic mineral whereas in the Great Dyke olivine is confined to the lowest exposed layers. The total ranges in feldspar composition are similar, although the Quihita-Cunene intrusion is richer in the earlier differentiates (range An_{80} to An_{50}) while the Great Dyke is richer in the later differentiates (range An_{76} to An_{44}).

The present paper is in the nature of a preliminary study, and it is hoped that it will serve to stimulate interest in, and further work on, this large intrusion. The hypothesis is advanced that it is a large, fractionated, layered basic intrusion, comparable in form and dimensions with the Great Dyke of Southern Rhodesia, and probably layered in the same manner (cf. Hess, 1950, fig. 3). Later investigators ought to consider this possibility, and to look for possible chilled contacts, layered structures (particularly dips of layering), later differentiates and possible extensions to the north of Chibia and to the south of the Rio Cunene.

Acknowledgments

The authors wish to thank Dr. G. R. McLachlan of the Port Elizabeth Museum for much help in the early stages of planning the expedition and geological programme, and Professor L. R. Wager, Dr. K. S. Sandford, and Professor R. W. Steel for help in both early planning and later working-up of the results.

For financial assistance to the expedition, thanks are due to the University of Oxford, the Alexander Allen Paton Memorial Fund, Balliol College, the Shell Petroleum Company, the Royal Geographical Society and the Geographical Magazine Trust, and to the Union Castle Line for financial concessions. Valuable help in the field was provided by Robert Hudson and Co. Ltd. and the Portuguese authorities.

References

Assunçao, C. T. de, F. Mouta, and J, Brak-Lamy, 1954. Contribution a l'etude de la petrographie du Sud de l'Angola. *XIX^e Cong. Geol. Internat., Alger 1952*, Fasc. xx, 253–273.

Beetz, P. F. W., 1933. Geology of South West Angola, between Cunene and Lunda axis. *Trans. Geol. Soc. S. Africa*, xxxvi, 137–176.

Brown, G. M., 1956. The layered ultrabasic rocks of Rhum, Inner Hebrides. *Phil. Trans. Roy. Soc.*, Ser. B, ccxl, 1–53.

Faber, F. J., 1926. Bijdrage tot de Geologie van Zuid-Angola (Afrika). *Proefschrift. Technische Hoogeschool te Delft.*

Mouta, F., and H. O'Donnell, 1933. *Carte Géologique de l'Angola* (1 : 2,000,000). Ministério das Colónias, Lisbon.

Souza, F. L. P. de, 1916. Contributions a l'etude Petrographique du Sud-ouest de l'Angola. *Compt. rend. Acad. Sci.*, clxii, 692–694.

VAGELER, P., 1920. Beobachtungen in Suedwestangola und im Ambolande. *Verh. Ges. f. Erdk.* (17.ix.1919), 179–193.

WAGER, L. R., 1953. Layered Intrusions. *Medd. Dansk. geol. Foren*, xii, 335–349.

—— and W. A. DEER, 1939. Geological investigations in East Greenland, Part III. The petrology of the Skaergaard intrusion, Kangerdlugssuaq. *Medd. om Grønland*, cv, No. 4, 1–352.

DEPARTMENT OF GEOLOGY AND MINERALOGY,
UNIVERSITY MUSEUM,
OXFORD.

EXPLANATION OF PLATE VI

FIG. 1.—Olivine-accumulate (A. 420) with poikilitic, interprecipitate, plagioclase. Typical of the ultrabasic rocks from close to the western margin. (× 55.)

FIG. 2.—Plagioclase feldspar accumulate (A. 352), with a few primary precipitate olivines modified by addition of olivine (narrow extensions) from the interprecipitate liquid. Typical of the basic rocks from the central parts of the intrusion. (× 15.)

FIG. 3.—Typical granite-gneiss country rock (A. 378), occurring both at the margins of, and as inclusions within, the gabbroic intrusion. Note the shadowy extinction of quartz, granulation at grain boundaries, and myrmekite (bottom right). X-nicols. (× 30.)

1

2

3

QUIHITA-CUNENE GABBROIC INTRUSION

Oblique-Slip Faults and Rotated Stress Systems

By Alwyn Williams

Abstract

Primary shears with oblique-slip displacement can only arise by failure after a rotation of two of the three principal axes of stress from an expected vertical and horizontal position relative to the earth's crust, and are either normal-wrench or thrust-wrench hybrids. As a consequence of a full 90° clockwise or anticlockwise rotation, the changes in pitch of striations on slickensided fault planes belonging to both categories are stereographically represented by two small circle courses which are $\angle 90° - \phi$ apart for the normal-wrench association, and $\angle 180° - \phi$ apart for the thrust-wrench one: $\angle \phi$ being the angle of internal friction. Using these assumptions to analyse the distribution of oblique-slip shears exposed in the south-western part of the Midland Valley of Scotland, it appears probable that two unrelated sets of commonly occurring normal-dextral shears are of primary origin, for the inferred azimuthal orientations of the principal axes of stress are remarkably close to those estimated for other, more orthodox tectonic features found elsewhere in the Midland Valley and appropriately absent in the south-west.

THE interpretation of fault systems according to the theory of stress as set forth by E. M. Anderson in his classic monograph on "The dynamics of faulting" (1951) has now become general practice. Failure of any brittle substance through stress differences, which are conventionally defined in terms of three principal axes, ideally results in shearing along two planes symmetrically disposed about the axis of maximum stress and subtending an angle of $90° - \phi$ where ϕ is approximately equivalent to the angle of internal friction of the material undergoing failure. All three fundamental fault types—wrench, normal, and thrust—may be explained in this way with the further limitation that one of the three principal axes of stress is an expression of gravity so that the other two axes, which are perpendicular to it and to each other, are stresses operating horizontally in the earth's crust. In these circumstances the manifest differences between the basic fault patterns are simply dependant on whether gravity is the maximum (σ_1), mean (σ_2), or minimum (σ_3) principal stress within any stress field. Normal faulting results when gravity is σ_1, so that the hade of a shear belonging to this system is $45° - \frac{\phi}{2}$; when gravity is σ_2, thrusts are produced with dips of $45° - \frac{\phi}{2}$ and when gravity is σ_3, wrench faulting occurs with the strikes of the fault planes deviating at $45° - \frac{\phi}{2}$ from the azimuth of σ_1.

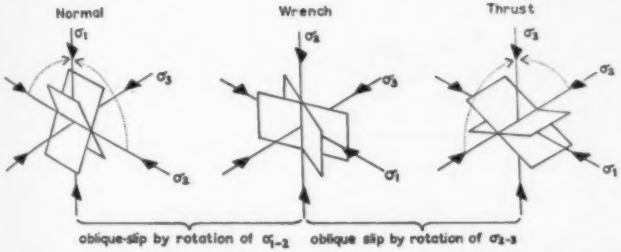
The theory is, of course, well substantiated by field data and the majority of faults can immediately be allocated to one of the three major categories, not only on the evidence of stratigraphical displacement but also, where exposed fault planes consist of slickensided surfaces, on the attitude of the striations. If the striations are parallel with the strike of a fault plane they were produced by a horizontal slip pivoted about a mean stress (σ_2) due to gravity; but if they run with the dip of a fault plane the stress due to gravity was either σ_1 or σ_3. In effect it is customary to think of wrenches as strike-slip faults, and normals or thrusts as dip-slip ones, although, since the hade of a normal fault, the dip of a thrust and the strike of a wrench fault are all equivalent to the $\angle 45° - \frac{\phi}{2}$, these radical differences are quite reconcilable in that the striational pitch is a function of $\angle 45° - \frac{\phi}{2}$ and so is related to the azimuth and inclination of the maximum principal stress.

Despite the theoretical implications that the striations of slickensided shears may be aligned in two directions only, viz. parallel with, or normal to the strike of a fault plane, oblique-slip fractures are widely known. For example, in the Girvan district, where more than 600 slickensided surfaces have been examined, over a quarter of these bear striations pitching at any angle intermediate between a strike-slip of 0° and a dip-slip of 90° and on commonness of occurrence alone they are sufficiently important to merit further consideration.

The assumption that the pitch of striations as a function of $\angle 45° - \frac{\phi}{2}$ is an estimate of the azimuth of σ_1, and its inclination, relative to the horizontal, leads to the inference that all obliquely pitching striations are the result of a maximum principal stress deviating from an expected vertical or horizontal direction by an amount proportional to the degree of pitch. Two explanations immediately suggest themselves. First, oblique-slip shear planes are due to stresses acting on material of anisotropic strength, e.g. the tilting or activation of fault planes which originated in an earlier unrelated stress field. Clearly, repeated slip along any fault plane with a long geological history will almost certainly reflect changing stress conditions in this way, although anisotropy poses such formidable problems in structural analysis that, at present, partial solutions may only be derived empirically. Alternatively, oblique-slip faults may have arisen as independent fracture systems and may result from stresses operating in the earth's crust which are not orientated with gravity as a major contributor to the dynamic stress system.

The principal axes of stress for the three standard fault types are

given in Text-fig. 1. An azimuthal swing of σ_2 in a normal stress field, and of σ_1 in a wrench or thrust field, will be accompanied by an equal rotation in the horizontal plane of σ_2, σ_3, and σ_3 respectively. Reorientations of this nature within one stress system have already been described by McKinstry (1953) and Moody and Hill (1956) and are responsible for the propagation of second- and third-order shears in wrench fault patterns. But all of them will necessarily be either strike-slip or dip-slip with a correspondingly orthodox alignment of striations, and the only possibility of producing oblique-slip shear planes lies in a swing from the vertical of σ_3, σ_1, and σ_3 of the wrench, normal, and thrust fields respectively, with a corresponding rotation of one of the horizontal axes. In each case deviation gives rise, on failure, to a hybridization of any two fault systems. Thus in both normal and

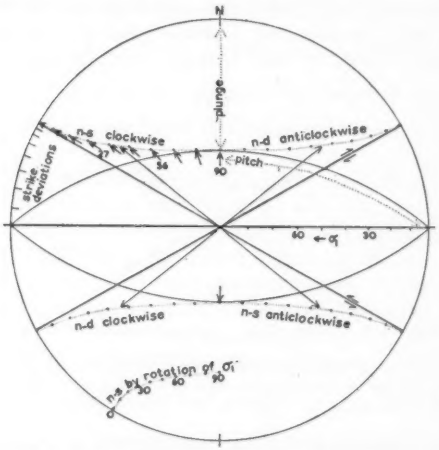


TEXT-FIG. 1.—The principal axes of stress for normal, wrench, and thrust fault systems, showing the rotations necessary to produce oblique-slip fractures.

thrust stress fields, failure after the rotation of the $\sigma_1 - \sigma_3$ axes through 45° will result in the appearance of a high angle reverse and a low angle normal. Hafner (1951, p. 394) has postulated that this anomalous fault pattern could occur in zones peripheral to areas undergoing differential uplift, and cites the Basin Ranges of Nevada as affording some confirmatory evidence of their existence. Such an association holds throughout an arc of $\angle\phi$ bisected by the 45° axis, but both sets of shears will be dip-slip. Indeed oblique-slip shearing can only occur through failure after planar rotation of the σ_1–σ_2 axes of the wrench-normal system and of the σ_2–σ_3 axes of the wrench-thrust association.

The minimum principal stress, σ_3, of both wrench and normal stress fields is horizontal, and a rotation of σ_1 and σ_2, through an arc of 90° about this axis transforms a normal stress system into a wrench one and vice versa. Thus, suppose that material, with an angle of internal friction of 30°, fractured under a normal stress system in which σ_2 was aligned north–south, two sets of dip-slip shears should occur striking east–west and hading at 30° north and south. The

complementary wrench shears should consist of vertical strike-slip planes striking at 60°–240°, and 120°–300°. For rotations of less than 90°, two resolutions of a stress field are possible dependent on whether σ_1 and σ_2 moved in a clockwise or anti-clockwise direction (Text-fig 1). Failure after a 45° clockwise rotation of σ_1 should produce shears hading at 20° towards north 22° E. and south 22° E. Striations on fault planes with northerly hades should plunge at 38°, west 39° north and pitch at 42° in the same direction and the net displacement consist of a horizontal sinistral component and a dip-slip normal one. Stria-



TEXT-FIG. 2.—The cyclographic projections of a normal and a wrench fault system with a coincident minimum principal stress (σ_3) aligned north–south and the traces of striational plunges of slickensided oblique-slip fractures, either normal-dextral (n-d) or normal-sinistral (n-s), resulting from a clockwise and anticlockwise rotation of σ_1 and σ_2.

tions on the southerly hading fault planes should plunge and pitch at 38° and 42° respectively 39° south of west; the total oblique-slip would then include a horizontal dextral component and a dip-slip normal one.

The two shear sets produced after a 45° anti-clockwise rotation of σ_1 from the vertical should have identical strikes with those appearing after a clockwise rotation of σ_1 through an equal arc, but they should hade in the opposite direction. Moreover the sense of rotation will be shown by slickensided surfaces because faults in this category will bear striations plunging at 38°, 39° north, and 39° south of east for the

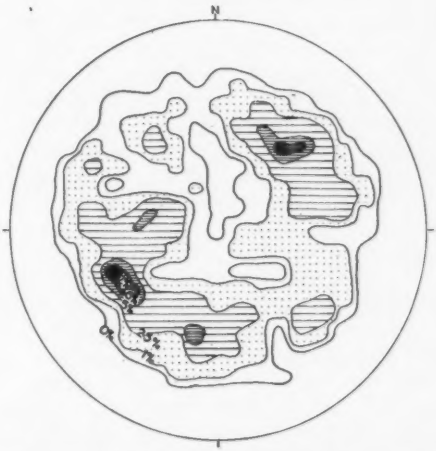
normal-dextral and normal-sinistral shears respectively, and pitching at 42° in the same directions.

All essential parameters of shear planes which are hybrids between wrench and normal faults may be calculated as follows:

$$\tan \alpha = \tan \left(45° - \frac{\phi}{2}\right) . \sin \gamma$$

$$\tan \beta = \tan \left(45° - \frac{\phi}{2}\right) . \cos \gamma$$

$$\tan \gamma^1 = \cos \left(45° - \frac{\phi}{2}\right) . \tan \gamma$$



TEXT-FIG. 3.—A stereogram showing the distribution of striational plunges which pitch at angles from 30° to 89° inclusive on 169 slickensided fractures, with hades less than 45°, affecting the Lower and Middle Palaeozoic rocks of the Girvan area.

where α is the hade, β, the azimuthal deviation from the strike of the normal fault, γ^1, the pitch of striations, and γ, the degree of rotation of σ_1 from the horizontal to the vertical. For purposes of structural analysis, however, such characteristics are more instructively represented on a stereographic net as in Text-fig. 2 where it can be seen that the plunge of the striations follow two small circle courses separated by an interval equivalent to $\angle 90° - \phi$.

Oblique-slip shears, which are hybrids between wrench and thrust faults and imply failure after the rotation of σ_2 and σ_3 about the horizontal σ_1 axis, may be analysed in a manner similar to that adopted

for the wrench-normal association. The hade, azimuthal deviation and pitch striation for any such hybrid plane developed during the rotation of σ_3 from the horizontal to the vertical, may be derived as follows:

$$\cos\ \alpha = \sin\left(45 - \frac{\phi}{2}\right) . \sin\gamma\ \mathrm{cosec}\ \gamma^1$$

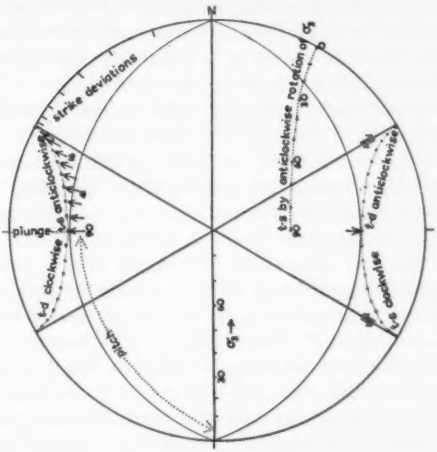
$$\tan\beta^1 = \tan\left(45 - \frac{\phi}{2}\right) . \sec\gamma$$

$$\tan\gamma^1 = \sin\left(45 - \frac{\phi}{2}\right) . \tan\gamma$$

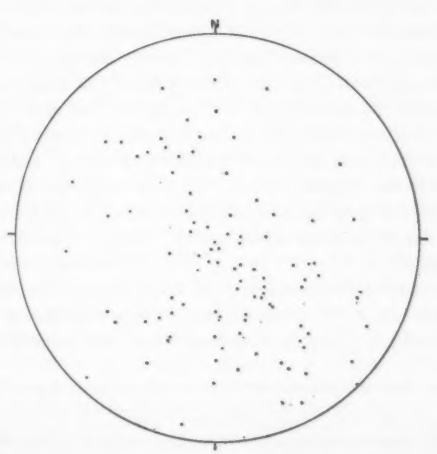
where $90 - \beta^1$ is the azimuthal deviation from the strike of the thrust. The attitudes of the plunge of striations are illustrated stereographically in Text-fig. 4, which shows that for deviations of σ_3 intermediate between 0° and 90°, two distinctive fracture sets are again possible, dependent on whether rotation was clockwise or anti-clockwise, i.e. northwards or southwards in the diagram. Consequently, for any one quarter-circle turn, the plunge of striations will follow two small circle arcs between an obtusely divergent strike-slip position along the wrench faults to a diametrically opposed dip-slip down the thrusts, and the completed traces for both senses of rotation will be two small circle courses separated by an interval equal to $\angle 180° - \phi$.

The practical application of these methods of analysis can be demonstrated by reference to the faulting affecting the Ordovician, Silurian, and Old Red Sandstone successions exposed in the south-west corner of the Midland Valley of Scotland, between Girvan and the Glen App segment of the Southern Upland Fault complex, some 10 miles to the south. The full tectonic history is somewhat involved and will be described elsewhere. Provisionally it may be outlined as consisting principally of a Caledonian phase of folding, thrusting, and wrench faulting, all resulting from a sustained maximum horizontal stress aligned at 160°–340°. These movements have not deformed either the Lower or Upper Old Red Sandstone successions which are tectonically complicated only by being affected by faulting, mainly of the oblique-slip kind.

The distribution of plunges of striations, which pitch at angles ranging from 30° to 89° inclusive on 169 slickensided fault planes, is given in Text-fig. 3. Two well-defined and almost diametrically opposed maxima occur for plunges at 45° to 50° to the north-east and south-west. Nearly a half of the 47 faults contributing to the double maxima are known from field evidence, such as stratigraphical displacement, tension gashes, etc., to have a normal-dextral displacement and there is no reason to suppose that the remainder of them are different. It is evident, however, that the maxima do not reflect failure



TEXT-FIG. 4.—The cyclographic projections of a thrust and a wrench fault system with a coincident maximum principal stress (σ_1) aligned east–west and the traces of striational plunges of slickensided oblique-slip fractures, either reverse-dextral (t-d) or reverse-sinistral (t-s), resulting from a clockwise and anticlockwise rotation of σ_2 and σ_3.



TEXT-FIG. 5.—The scatter of striational plunges of 91 dip-slip reverse fractures affecting the Lower Palaeozoic rocks of the Girvan area.

after clockwise or anti-clockwise oscillations of the σ_1–σ_2 axes in the same stress field. If this were true, the average strike of the faults in both clusters would be the same, whereas they are significantly different and so necessitate independent interpretation.

The twenty-two faults belonging to the north-east maximum have a mean strike of 48° and dip at 70° or more to the north-west. Now the angle of internal friction can vary considerably but Anderson (1951, p. 56) has concluded that a value of $\angle\phi = 45°$ may be taken as representative for most conditions of fracture in sediments and Hubbert (1951, p. 362) has suggested 30°–35°. In the Girvan district a mean $\angle\phi$ of 30° has been determined empirically from a study of fully developed shear systems in 14 separate Lower Palaeozoic localities, and is sufficiently close to those given by other authors to be acceptable for the present analysis.* The plunges of striations for any oblique slip faults will therefore follow two small circle traces, separated by a great circle arc equivalent to $\angle 90 - \phi$ as in Text-fig. 2, and from this diagram it is possible to calculate that normal-dextral faults, with a strike of 48°–228° and a striational plunge of 50°, belong to an association of a normal fault striking at 72°–252° and hading northwards at 30° and a vertical dextral wrench striking at 42°–222°.

From these data it is possible to conclude that the stress field, giving rise to the oblique-slip faults under discussion, was orientated with a horizontal σ_3 aligned at 162°–342°, a σ_1 inclined at 50° towards east 18° north, and a σ_2 inclined at 40° to the west 18° south (Text-fig. 6A).

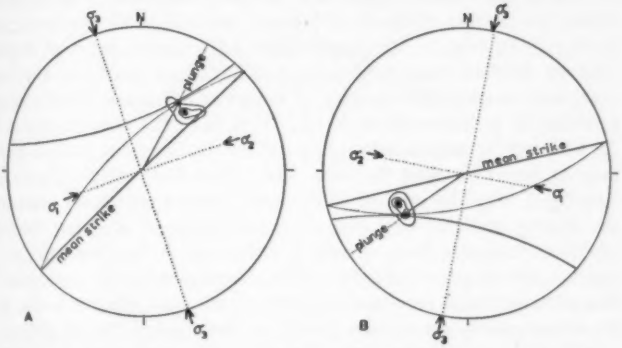
The mean strike for the twenty-five normal-dextral faults comprising the south-west maximum is 77°. By adopting the same procedure used to analyse the origin of the north-east maximum, it is possible to interpret this group of faults as resulting from failure in a stress field with a horizontal σ_3 trending 11°–191°, a σ_1 inclined at 50° to the west 11° north and a σ_2 inclined at 40° to the east 11° south (Text-fig. 6B).

The reconstructed stress fields for the two sets of normal-dextral faults found in the Girvan area appear then to be unrelated. Those striking at 42°–222° are due to a relief of pressure in a 162°–342° direction (group A), while those striking at 77° have a minimum principal stress orientated at 11°–191° (group B). Nevertheless they fit very neatly into Anderson's chronology of stress changes for the Scottish region and provide a striking confirmation of his synthesis.

The stress field of group A is reminiscent of that existing during the

* The slight discrepancies between the small circles traces of striational plunges and the maxima for oblique-slip shears as revealed in Text-fig. 6 can be eliminated if $\angle\phi$ is assumed to be 40°. This value is very much closer to that used by Anderson to analyse fractures in similar media but for purposes of comparison with the theoretically derived results of Text-fig. 2, the Lower Palaeozoic value of $\angle\phi = 30°$ is retained, without, of course, affecting the positions of the principal axes of stress.

later phases of the Caledonian orogeny, in that σ_3 of the former and σ_1 of the latter coincide. Text-fig. 7 gives the distribution of plunges of striations, pitching at less than 29°, which are preserved on 348 shear planes. These and others, which did not shear with slickensided surfaces, represent the pre-Old Red Sandstone wrench system and from a study of them it has been possible to obtain over 200 estimates of the position of the maximum principal stress, σ_1, which show a marked mode for a 160°–340° alignment. Anderson (p. 180 *et seq.*) has



TEXT-FIG. 6.—The inferred stress systems for the fractures contributing to the two maxima of oblique striational pitch shown in Text-fig. 3.

pointed out that two phases of dyke intrusion are known with this orientation for σ_3, firstly the Etive and Ben Nevis swarms, which are, however, pre-Middle Old Red Sandstone in age, and secondly a series known mainly in the western part of the Midland Valley, " connected with the Calciferous Sandstone episode of vulcanicity " (p. 183). Group A faults therefore are quite likely to have originated during this period and so belong to Anderson's Proto-Armorican stress phase.

The stress field proposed for the faults of group B is also in harmony with that envisaged for other important tectonic elements in Scotland. A conspicuous series of east–west trending quartz-dolerite dykes and normal faults which affect the Carboniferous sediments of the Midland Valley have been analysed by Anderson (p. 29 *et seq.*). He concluded that both intrusion and shearing owed their origin to a maximum relief of pressure aligned north–south and that this crustal condition, the Borcovician stress system, persisted from late Carboniferous to Middle Permian times. Anderson (p. 33) provides a histogram of the trend of those Borcovician faults (along with Tertiary ones) known in the Midland Valley: their secondary modal strike is 100°–280°, so that the minimum principal stress axis, σ_3, was aligned in the 10°–190° direction

which is for all practical purposes coincident with the 11°–191° orientation for the group B faults, and it is probable that they too originated in late Carboniferous-early Permian times.

The importance of the thrust-wrench associations as contributors to the data on oblique-slip faults needs no more than a note. It has already been stated that thrusting in the Girvan area was limited to the Caledonian phase of deformation which also included episodes of folding and wrench faulting, all confirming the persistence of an orthodox stress field with σ_1 aligned in a 160°–340° direction. The distribution of the plunges of dip-slip striations preserved on 91 reverse shears is given in Text-fig. 5. The plunges show a fair scatter, although 80 per cent are confined to the north-west and south-east quadrants and the maximum concentration consists of a number of reverse faults hading at about 30° to the south-south-east. This, of course, is to be expected for they are the second-order splays off the far fewer but more important thrusts which affect the area. The wrench fault systems shown in Text-fig. 7 were the product of the same general maximum stress as the folding and thrusting and, since the plunge of striation for all phases of transition from wrench to thrust will be less than 30°, any significant development of hybrid faults should show in this stereogram. The unusually wide compass range of wrench fault maxima is due to the occurrence in the Girvan district of two distinct sets of sinistral and dextral shears representing a first- and second-order series. If either set had been involved in a shearing complex resulting from a rotation of σ_2 and σ_3, one would expect to find a concentration of plunges somewhere along one of the four small circle traces on the stereogram. One such cluster does occur around a plunge of 15° towards north 21° west, and if it does represent a 30° clockwise rotation of σ_3 from the horizontal, the faults composing it will be reverse-dextral ones striking at 152°–332° and hading eastwards at 25°. Only 6 out of the 14 faults involved fulfil these requirements: the remainder have either negligible hades with corresponding discrepancies in strike or are even inclined westwards, suggesting a normal component to the oblique-slip in place of a reverse one, and the data are not considered as providing any conclusive evidence that a rotation of σ_2–σ_3 ever took place.

The use of rotated stress systems to interpret oblique-slip faults will probably cause some concern, and it must be emphasized that the writer by no means advocates the approach outlined above as affording a general solution to the problem of oblique-slip faults. Quite obviously all other possible complications have first to be eliminated and indeed it is for this reason that inclined axes of stress have been used to explain some of the oblique-slip faults in the Girvan area. Thus the possibility of attributing all observed pitching striations to

the effects of tilt seems to be remote, for, if striations on post-Caledonian faults have been tilted out of original vertical or horizontal attitudes of pitch, the older thrust and wrench systems would show commensurate deviations, whereas the vast majority do not. Anisotropic effects are admittedly less decisively rejected for the simple reason that they cannot easily be assessed but on the whole it appears unlikely that the normal-dextral fault sets represent the re-activation of pre-existing fractures. Their strikes do not correspond with any of the maxima for the wrench

TEXT-FIG. 7.—A stereogram showing the distribution of striational plunges which pitch at angles from 0° to 29° inclusive on 348 slickensided wrench fractures affecting the Lower Palaeozoic rocks of the Girvan area.

systems and even if it is conceded that the faults of group A were originally sinistral wrenches belonging to the 35°–215° maximum of Text-fig. 7 and that those of group B were re-activated members of the 85°–265° dextral peak, the pronounced clusters around the 45° plunges which characterize the normal-dextrals would more likely continue as arcuate trails towards their supposed parent wrench maxima. Moreover, with regard to the faults of group B it is probably highly significant that the east–west normal faults system, so prevalent in the Midland Valley, is not known in the Girvan area. Certainly the stress conditions which gave rise to such a pattern are known to have been quite widespread throughout Scotland and N. England and since the inferred stress orientations, apart from axial inclinations, for the faults of group B concord so closely with them, it seems plausible to conclude

that the normal-dextrals of Girvan are the local equivalents of the Borcovician normal fractures developed elsewhere.

Hafner (1951) in his discussion (p. 391 *et seq.*) on the effects of vertical stresses superposed on those of the standard state points out that such vertical stresses must die out laterally and has assumed that this variability is most conveniently expressed as a sinusoidal curve which " also forms the most satisfactory approximation to many actual cases of differential vertical uplift " (p. 391). The resultant stress system and the expected shear pattern (pl.1, fig. A) are constructed in the plane of the minimum and maximum principal stresses and show that in zones peripheral to a positive wave-length of vertical stress variation, σ_1 and σ_3 may rotate out of their respective vertical and horizontal positions by as much as 45°. The corresponding shears would consist of dip-slip high-angle reverses and low-angle normals. In a like manner no vertical stress can be sustained indefinitely along the axes of mean principal stress, and if its variation in this direction were also sinusoidal, oblique-slip normal-wrench faults similar to the ones described here would result.

I am greatly indebted to Dr. I. Adamson, Mr. T. B. Anderson, and Dr. W. Schwarzacher of Queen's University, Belfast, and Dr. D. A. Bassett of Glasgow University for their helpful advice on various aspects of the manuscript.

REFERENCES

ANDERSON, E. M., 1951. The dynamics of faulting. Oliver and Boyd Ltd., London.

HAFNER, W., 1951. Stress distribution and faulting. *Geol. Soc. America Bull.*, lxii, 373–398.

HUBBERT, M. K., 1951. Mechanical basis for certain familiar geologic structures. *Geol. Soc. America Bull.*, lxii, 355–372.

MCKINSTRY, H. E., 1953. Shears of the second order. *Am. Jour. Sci.*, ccli, 401–414.

MOODY, J. D., and M. J. HILL, 1956. Wrench-fault tectonics. *Geol. Soc. America Bull.*, lxvii, 1207–1246.

DEPARTMENT OF GEOLOGY,
THE QUEEN'S UNIVERSITY,
BELFAST.

The Geology of the Land between Ekmanfjorden and Dicksonfjorden in Central Vestspitsbergen

By D. E. B. Bates and W. Schwarzacher

Abstract

The peninsula between Ekmanfjorden and Dicksonfjorden has been mapped geologically. Devonian sandstones are overlain unconformably by a Permo-Carboniferous sequence of limestones (200 m.) and dolomites and gypsum (300 m.). This is followed by more than 150 metres of Permian cherts. The strata show a slight but consistent dip towards the SSE. A fault having the trend of this dip has been proved in Ekmanfjorden. Sills are intruded in the higher levels of the succession and it is suggested that the intrusions are contemporaneous with the faulting. By comparing the lithology the area can be correlated with the known sections of Billefjorden and Kongsfjorden.

Introduction

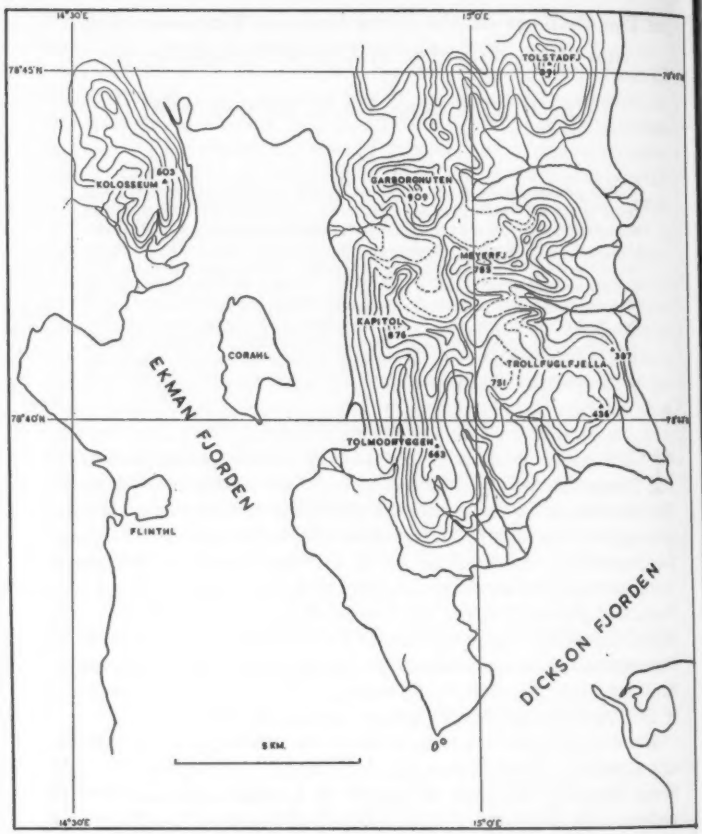
THIS paper deals with the geology of the peninsula between Ekmanfjorden and Dicksonfjorden in the north part of Isfjorden. The area has not been described in detail before and in fact only a few geologists have visited it. Nordenskiöld (1876) figured a section through the peninsula from south to north, in which Devonian Sandstones, Carboniferous limestone, and dolerite sill intrusions are differentiated. Føyn and Heintz (1943) of the English-Norwegian-Swedish Paleontological expedition visited the head of Ekmanfjorden to collect from the Devonian sandstones. Orvin (1940) in compiling a geological map of Svalbard made use of aerial photographs to work out the boundaries of the Permo-Carboniferous systems against the Devonian.

The two authors spent three weeks on the peninsula during a university expedition from Belfast to Spitsbergen in the summer of 1957. Some mapping was done on a scale of 1:50,000, and a number of sections were measured using a 6-inch Paulin aneroid together with a 3-inch survey aneroid and in some instances using an artillery director for triangulation. Further details were taken from ground and aerial photographs (Text-figs. 1 and 2).

The Sedimentary Rocks

The Devonian Sandstones are exposed only in the north of the area where the succession is as much as 300 m. thick and according to Føyn and Heinz (1943) belongs to the Lykta and Kapp Kjeldson divisions of the Wood Bay Series.

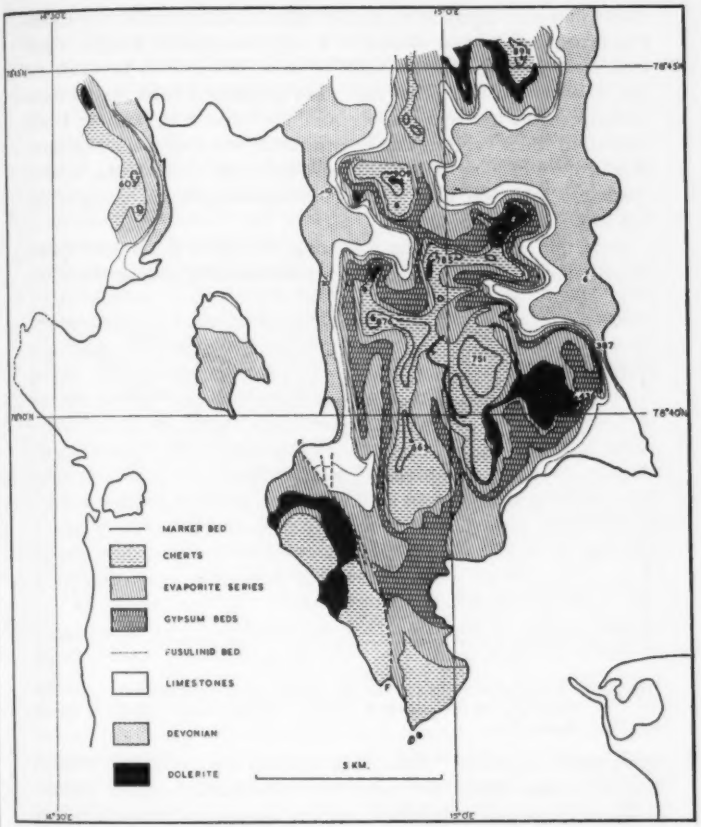
The outcrops examined on the eastern shores of Ekmanfjorden were of a massive dark red-brown, micaceous sandstone, which is commonly current bedded, associated with occasional thin lenses or beds of red



TEXT-FIG. 1.—Topographical map of the peninsula between Ekmanfjorden and Dicksonfjorden, compiled from Isachsen (1915), and the Norsk Polarinstitutt chart 503, supplemented by a plane table survey.

shale and a fine grey-green conglomerate. Both sandstone and conglomerate have a matrix which is partly ferruginous and partly calcareous; the sand grains are of quartz, while pebbles of fine-grained limestone are common in the conglomerates.

The Permo-Carboniferous. The lowest member of the Carboniferous succession is a well-developed limestone with an average thickness of 200 m. This limestone rests with a basal conglomerate on the Devonian Sandstone filling in slight irregularities of the pre-Carboniferous sur-



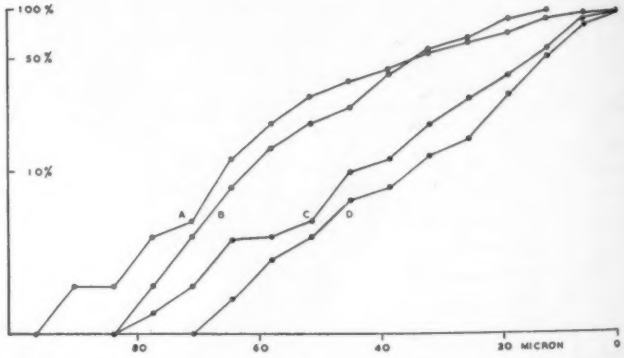
TEXT-FIG. 2.—The geology of the area shown in Text-fig. 1.

face. The basal conglomerate has been seen *in situ* only in one locality where it is a few centimetres thick, and it is probably reasonable to assume that it is nowhere very much thicker. The pebbles in the conglomerate are small, varying from 5 to 10 mm. in diameter, and consist mainly of Devonian quartz sandstone and older quartzites embedded in a reddish sandy matrix.

The limestone itself is light grey to white in colour. In the lower two-thirds it forms steep cliffs, in the upper third it weathers to gentler slopes. Examination of the limestone in polished and thin sections

shows that it is largely formed from organic debris, foraminifera, shell fragments, and crinoid ossicles in a very fine-grained matrix. Commonly the cement has recrystallized as a clear calcite in which the micro-fossils are well preserved. The calcareous mud matrix occasionally shows a slight grading and the fossil fragments in it are orientated which suggests that its deposition was a mechanical process. Sedimentary bedding on a scale sufficiently fine to be found in hand specimens is usually absent, but contemporaneous reworking of the limestone has been observed.

Most thin sections of the limestones contain 1–5 per cent quartz grains, mainly of clastic origin, though occasionally idiomorphic authi-



TEXT-FIG. 3.—Cumulative curves of typical size distributions from the limestone series (curves C and D) and the evaporite series (curves A and B).

genic quartz is present. The clastic quartzes are mostly well rounded and often show signs of corrosion due to diagenetic processes. Microscopic size analyses of insoluble residues show that the size frequency curves follow a logarithmic percentage law whereby the means of the grain sizes vary between 20 and 70 micron (Text-fig. 3). Doeglass (1956) finds similar distributions in sediments which have been directly deposited from suspensions.

The limestone succession frequently contains layers of nodular chert, usually grey to black in colour. Many of the nodules contain a pronounced concentric banding ; the chert always replaces the limestone with diffuse boundaries to the silicified areas. The fact, however, that the chert nodules occur in horizons strictly parallel to the sedimentary bedding suggests that their stratal disposition has been somehow controlled by primary sedimentary concentrations. In several chert specimens small traces of fluorspar occur in cracks in the chert closely

associated with calcite. The crystals are dark blue but hardly ever bigger than 1 mm.

Dolomite is relatively rare in the specimens of the limestone sequence but there are several bands which are strongly dolomitic. The bands can be recognized in the cliff section by their white colour and their tendency to weather more easily, they are only 20 to 30 cm. thick and they occur near the more strongly developed bedding planes of the limestones. Several of these bands have been traced throughout the entire area and there is little doubt that the concentration of magnesium in the bands is a primary one.

The most conspicuous marker horizon occurs in the upper third of the limestones. It is a thick band of a dark bituminous limestone, approximately 3 m. thick, containing innumerable fusulinids embedded in a fine-grained calcilutite and with dark films of bitumen between the grains. In one specimen which is taken from the top of the fusulinid bed the matrix is replaced by small idiomorphic dolomite crystals.

The bedding in the succession is well developed with prominent bedding planes spaced at an average thickness of 19 m. The thick bands caused by them are visible from a great distance and often give the slopes a stepped appearance. Such " master " bedding planes may be traced without difficulty throughout the area (Text-fig. 4) either by following an individual bed in the cliff sections directly or by using the main lithological boundaries as reference levels for individual bands (cycles). Apart from the " master " bedding planes bedding also occurs on a smaller scale to differentiate beds 2–5 m. thick. These bedding planes are less persistent in that they may disappear laterally only to reappear at the same horizon in a different locality. It is evident that the pattern of the sedimentary bedding is caused by fluctuations of the factors controlling sedimentation. Not enough data are available to state exactly which lithological factors produced the bedding planes, although it has been observed that most of the dolomite bands and the chert layers form close to the prominent master bedding planes. What has been established is the constancy of the master bedding planes and therefore their usefulness for detailed stratigraphical correlation within the area mapped. Eight such bands (cycles) have been recognized within the limestone succession but as can be seen from Text-fig. 4 the lower two may be missing due to unconformable overlap. The fusulinid band occurs on top of cycle 5. The highest cycle, 8, is usually more thinly developed and shows, on its top, signs of reworking such as cross-bedding and channelling and in one locality a loose specimen of conglomerate has been found.

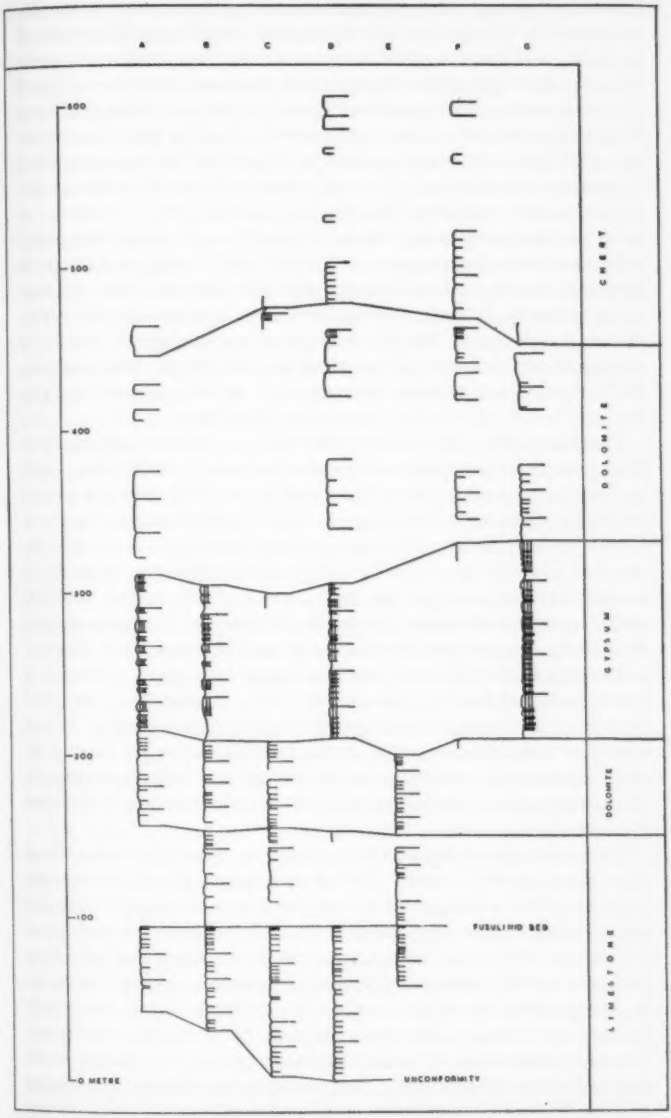
Few data are available to reconstruct the conditions under which the limestone has formed. The transgression took place over fairly flat ground and limestone deposition started almost immediately. The

sea was sufficiently shallow to allow reworking of the sediments but no erosional surfaces within the limestone have been observed. The occasional bands of dolomite are best explained by changes in the salinity of the sea which may indirectly be responsible for the sudden occurrence of masses of fusulinids. Land was sufficiently close to supply a steady influx of clastic material. Towards the end of the limestone series deposition slowed down and parts of the highest cycle became eroded. This break need not be of long duration but it clearly divides the limestones from the following evaporite series.

The Evaporite Series.—This consists of a succession of about 300 m. of dolomites, limestones and variable amounts of gypsum and anhydrite beds. Almost throughout the whole area a three-fold division of the evaporite series can be made out. The lowest member is a dolomite-limestone group, 50–60 m. thick, in which gypsum and anhydrite occur rarely in thin bands and veins. This sequence is followed by 90–100 m. of well-bedded anhydrite and gypsum and the series ends with 150 m. of dolomites and limestones. The lower dolomite group always forms steep cliffs or prominent crags weathering to a brownish rubble. The bedding is slightly less regular than in the limestones, but again prominent master bedding planes can be made out forming four beds with an average thickness of 14 m. These beds are further subdivided into less prominently bedded units, 0·5–1 m. thick, and in many cases there also exists a fine lamination which can give the rocks a shale-like appearance. From 50 measurements in hand specimens the average thickness of the lamination has been determined to be 4·3 mm. with a variation coefficient of 38 per cent. Bedding on this small scale is due to alternations of layers rich in detrital quartz with others formed of fine-grained dolomite. The detrital quartz in such a bed occurs in the lower half of the lamina, sometimes showing a slight grading towards the top. The boundary with the dolomitic lamina however is sharp and commonly shows slight irregularities which may indicate small-scale erosion before the deposition of the dolomite; in one case even cross-bedding on a minute scale was observed within the sandy layer. The dolomite seems to be recrystallized as it consists of a fine-grained mass of quite well-defined rhombohedra. The division into a lower sandy part and upper dolomitic part of the lamina is not always very clear as thin sandy streaks may also occur in the dolomitic portion of the

TEXT-FIG. 4.—Correlated sections from A: Kolosseum, B: South side of Garborgnuten, C: North end of Kapitol, D: summit of Kapitol, E: South end of Kapitol ridge, F: West side of Tolmodryggen, G: South side of Trollfuglfjella.

On the diagram strongly developed bedding planes are drawn longer than minor bedding planes. Vertical shading indicates beds of gypsum and anhydrite. The gaps in the sections are either not surveyed in detail or scree covered.


A
B
C
D
E
F
G
600
500
400
300
200
100
0 METRE
CHERT
DOLOMITE
GYPSUM
DOLOMITE
LIMESTONE
FUSULINID BED
UNCONFORMITY

lamina. In several examples it has been found that the top of the dolomitic part is very porous and sometimes contains small amounts of gypsum. Size analysis of the clastic material gives distribution curves (Text-fig. 3) of the truncated type which Doeglass (1956) found typical for salt marshes. Such size distributions, which are strikingly different from the one found in the limestone series, indicate that the sediment has lost some of its coarse material which may be due to repeated deposition and erosion. Pure limestones are rare or absent in this succession but occasional fine-grained limestones with admixtures of dolomite have been found. In such cases it can be seen that isolated dolomite rhombohedra grow within the calcite mud, and one finds transitional stages between almost pure dolomite and lime rich sediments whereby dolomite replaces limestone metasomatically. It is, however, noteworthy that the boundaries between higher and lower concentrations of dolomite never cut bedding planes, which suggests that original concentration differences of the magnesium are again responsible for this kind of compositional bedding.

Specimens with well-developed lamination sometimes contain structures interpreted as ripple marks. In sections the stratification runs in very regular sinuoidal folds with an amplitude of 3–8 mm. and a wavelength of 35–40 mm. Such ripples occur between undisturbed horizontal laminae, and their internal structure and shape is always symmetrical, although the tops of the ripples have sometimes been slightly eroded. In one example the irregularities of the surface probably caused by ripple formation has led to slumping on a very small scale. Possible mudcracks have also been observed in a 3 mm. thick siltband.

Laminated dolomites may pass vertically into sandy dolomites in which traces of bedding are absent. This change is accompanied by a loss of orientation by the quartz grains, which are always aligned with their longest axes parallel to the bedding of the laminated dolomites. Moreover, a specimen of the transitional zone shows that the lamination so characteristic of the dolomites was increasingly disturbed by reworking.

The conditions under which the dolomite series was formed must have been relatively stable. The sedimentation fluctuated regularly depositing clastic material which may have been eroded and deposited several times. Before deposition of the dolomite started a short pause is indicated within the lamina, followed by increased evaporation as indicated by the presence of precipitated gypsum in some of the cycles. It is impossible to decide whether the dolomitic layers are strictly primary or secondary although even in the latter case the transformation into dolomite would be dictated by the primary distribution of the magnesium in the sediment. Small-scale cross-bedding, symmetrical ripple marks, and somewhat doubtful mud cracks would suggest a very

quiet shallow sea, but little can be said about the actual depth. The control of the lamination is most likely to be climatic and may be annual. Occasionally the quietness of the sedimentation has been disturbed and the lamination obliterated by reworking. The rarity or absence of fossils supports the idea of a sea with a high salinity.

The succeeding gypsum-anhydrite beds show a certain facies variation. They are best developed in the south along the southern ridge of Kapitol and in the Trollfuglfjella group. There the evaporites are well stratified with beds of pure gypsum and anhydrite, 4–10 m. thick, separated by thin layers of dolomite which are sometimes bituminous. Master bedding planes are very well developed by the interspersion of thicker dolomitic beds which give rise to a large-scale stratification with beds of approximately 20 m. thick. The gypsum-anhydrite group consists of five such beds which can be traced laterally over the whole area, and by using them for correlation it was found that the dolomites between the gypsum beds increase in thickness at the expense of the latter towards the north of the area.

North of Garborgnuten and Tolstadfjellet the evaporite series loses its threefold division into lower dolomites, evaporites, and upper dolomites. A similar thinning of the gypsum beds can also be seen on the west side of the fjord on Kolosseum, but there is no indication that the post-Carboniferous fault which runs along the fjord affects the facies distribution of the evaporites.

The petrology of the evaporites has not been studied in detail. A specimen which has been sectioned shows a layer of anhydrite nodules in a rock which is approximately 50 per cent anhydrite and 50 per cent gypsum, simulating a gneissose structure with anhydrite augen. Most of the gypsum seems to have been formed after anhydrite.

The topmost 150 m. of the evaporite series is again dominated by dolomites though the middle part of this succession is almost invariably obscured by scree. The upper dolomites consist of three lower cycles defined by well-developed master bedding planes, the scree-covered section which is a shaly dolomite with minor gypsum bands, and finally two massive dolomite cycles ending with a band of gypsum about 5 m. thick. The lithology of the upper dolomites is very similar to that described for the lower dolomites.

Cherts.—The evaporite series is followed by a thickness of over 150 m. of chert, which are the youngest sedimentary rocks of the area. The base of this chert always forms a pronounced cliff, a feature which can be easily traced. The lowest chert band is about 20 m. thick and consists of dark grey chert beds, 2 to 3 m. thick. These are separated from one another by thin limestones which are crowded with fossils, mainly spiriferid brachiopods. These shell accumulations indicate an increased current activity, for the skeletal remains are commonly frag-

mented and bear signs of abrasion. Further they are interleaved with thin conglomeratic beds, with pebbles up to 0·5 cm. diameter, occurring between the more massive chert beds. Large portions of the chert in the lower parts of the succession have probably been formed by the silicification of a limestone. Various stages in this replacement can be seen and at times the lowest 20 m. of the chert series is in fact a limestone in which only the large shells have been silicified. The higher parts of the chert series are formed of cherts with occasional limestones for the most part obscured by a fine-grained scree. About 150 m. from the base of this series two very prominent bands of a dark grey chert occur and, since they can be traced as marker horizons over the whole area, the lowest of the two bands has been mapped.

There can be little doubt that the base of the chert series marks a renewed transgression into the area. The sea was presumably shallow at first as indicated by the numerous recurring shell bands.

Structure and Igneous Intrusions

The dip of the Permo-Carboniferous succession, 2° to 4° to the SSE, is in accordance with the regional dip around Isfjorden as seen on Dicksonland and in the Billefjorden area. This dip also affects the unconformity between the Devonian and Permo-Carboniferous strata which in the north has a height of 300 m. and reaches sea-level halfway to the southern end of the peninsula. From distant observations and aerial photographs it can be seen that the unconformity rises steadily towards the north of the area until the younger beds are restricted to the mountain tops. The red sandstones below the unconformity are not parallel to the Permo-Carboniferous, and dips up to 20° have been measured.

The Permo-Carboniferous is displaced along Ekmanfjorden by a normal fault trending NNW–SSE. The fault runs east of Coraholmen into the south-western part of the peninsula. It can be traced along the eastern boundary of the sill which forms the low ground west of Tolmodryggen. The throw of this fault is about 200 m.; fault drag leads locally to steeper dips in the country rock. The fault is subsequent to the sill intrusion on the south-west side of the peninsula. The NNW–SSE trend of the fault is parallel to the block-faulting of the more northern Devonian Sandstones. This might suggest that the fracture follows earlier lines of weakness similar to the faults in the Billefjorden area (see McWhae, 1952).

The outcrops of the dolerite sills are more widespread than indicated on the previous maps of Tyrrell and Sandford (1933) and Orvin (1940). It is seen that the stratigraphical level of the sill intrusions is almost constant throughout the peninsula, near the boundary of the evaporite series and the cherts. The large sill near Anservika in the Bille-

fjorden area (Gee, Harland, and McWhae, 1952) occurs at the same level. Further hitherto unrecorded sills have been observed on a reconnaissance sledge trip from Ekmanfjorden towards the west at Kongsvegsata and Palatium in similar positions.

Petrological examination of the dolerite shows a rock which varies from a fine-grained dolerite to a coarse-grained gabbro. The mineralogy compares closely to that described by Tyrrell and Sandford (1933). Olivine is absent and the principal dark mineral is pyroxene. Tyrell describes two types of pyroxene in his paper: one a pale brown variety with a low optic angle approximating to the enstatite-augite or pigeonite varieties; the other a colourless almost uniaxial pigeonite. In these rocks the pyroxenes are all pale brown, but with very small optic angles and with a $Z\Lambda C$ angle of 28°—suggesting that they are pigeonites with a composition of approximately 80 per cent clino-enstatite, and 20 per cent diopside. Hour-glass structures may indicate the presence of some titanium.

Plagioclase is found in idiomorphic laths, with an average composition of An60, but with zoning from a more anorthitic core to a rim of composition An30. Magnetite and ilmenite are common. In the mesostasis, chlorite and biotite are present and show pleochroic haloes round tiny grains of sphene. Intergrowths of feldspar and quartz are interesting since they seem to indicate a eutectic crystallization of a residual liquid rich in silica. Some serpentine suggests that olivine may have been present.

A Rosiwal analysis shows the following volumetric percentage composition: Magnetite, 7·47; Plagioclase, 48·1; Pyroxene, 37·8; Quartz, 1·53; Mesostasis, 3·06.

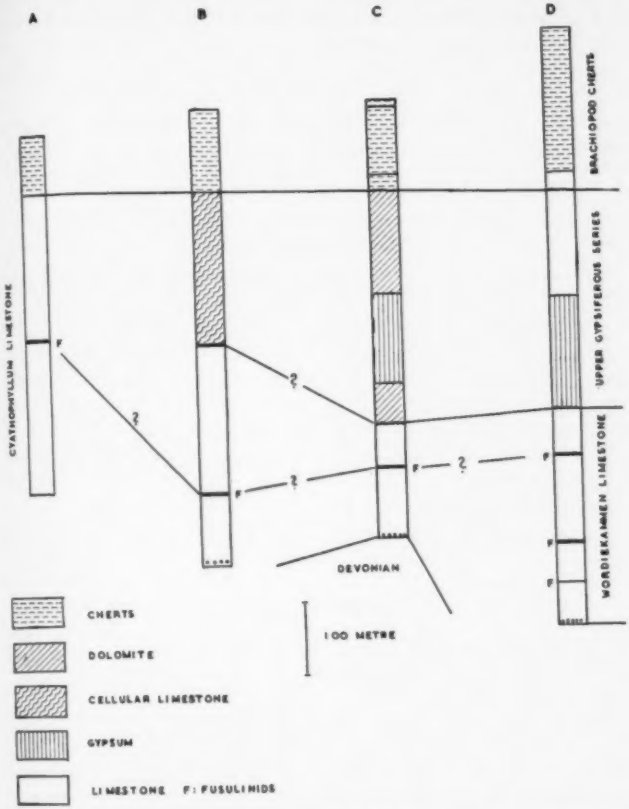
Although the sills occur at a nearly constant stratigraphical level they are not continuous over large areas. The sill in the south-west of the peninsula is partly intruded into the upper dolomites and partly into the lower members of the chert series. An apparent transgression of this sill in the south may be the effect of a fault which could have escaped mapping due to lack of exposures. Although the sill has an exposed thickness of 100 m. it does not occur at the corresponding horizon east of the fault. A second large sill is exposed in the upper dolomites of Trollfuglfjella. This sill is strongly transgressive along a direction NNW parallel to the valley between Tolmodryggen and Trolfuglfjella, where it passes into a vertical dyke. The country rock shows slight displacements along this line but throws are very variable and the sill seems to disappear towards the north. There is also a group of isolated dolerite exposures on Meyerfjellet, Kapitol, and Tolstadfjellet, which all occur between the evaporite series and the cherts. On Garborgnuten smaller sills were intruded into two stratigraphical levels, viz. the lower dolomites and fairly high up in the cherts.

From the field evidence it seems likely that the intrusion of the sills has been influenced by the structural direction NNW and that the faulting either precedes the intrusion slightly or is contemporaneous with it. From the geology of the Isfjorden area it is known that faulting occurs during Tertiary time, and it seems likely, therefore, that intrusion took place at the same time.

Correlation

As pointed out previously, it is possible to correlate a great number of lithological horizons throughout the area and it is likely that this detailed correlation can be considerably extended, but for the present only the major lithological groups can be recognized in neighbouring areas. A detailed stratigraphical study exists for the Billefjorden area (Gee, Harland, and McWhae, 1952) towards the east of Ekmanfjorden, and also for the Kongsfjorden region (Holtedahl, 1912; Orvin, 1934) on the west coast. Both areas show some development of middle and lower Carboniferous sequences which are missing from Ekmanfjorden. The lowest strata in Ekmanfjorden are the Cyathophyllum Limestones. This group has been subdivided in the Billefjorden area into the Wordiekammen Limestones, corresponding to our limestone series, and the Upper Gypsiferous Series, corresponding to our evaporite series. The cherts on top are given the name Brachiopod Cherts, following the nomenclature of Gee, Harland, and McWhae. The correlation is based on the coincidence of two small disconformities at the base and at the top of the gypsiferous series in both areas. Gee, Harland, and McWhae used two marker horizons for mapping these boundaries: the top of the Wordiekammen Limestones (limestone B), and the Spirifer Limestone at the base of the Brachiopod Cherts which has been identified in several parts of Spitsbergen. The second horizon can easily be recognized in our section but is usually more silicified than in the Billefjorden area. The boundary between the Upper Gypsiferous Series and the Wordiekammen Limestones is more doubtful. The top of the limestones in the Ekmanfjorden area never forms steep cliffs as in the Billefjorden area and the sharp change from limestone to dolomite has not been recorded in the Billefjorden area. On the other hand at the same distance below the disconformities in both areas occurs a black band of " Fusulina " limestone, which at least in the Ekmanfjorden area is a very persistent horizon. It seems therefore that the correlation shown in Text-fig. 5 is the most likely one, which means that the lower parts of the Wordiekammen Limestone are missing in the Ekmanfjorden area, a possibility which is borne out by the observed overlap of the limestones on to the Devonian Sandstones.

Correlation with more westerly sections as given by Holtedahl (1913) for the Dronningfjella and the Kongsfjorden area depends on only one



TEXT-FIG. 5.—The Permo-Carboniferous of A: Kongsfjorden (after Holtedahl, 1913), B: Dronningfjella (after Holtedahl, 1913), C: Ekmanfjorden, D: Billefjorden (Gee, Harland, McWhae 1953).

boundary, that between the cherts and the Cyathophyllum Limestone. Holtedahl has been unable to find the Spirifer Limestone and suggests that this formation has thinned out before reaching the Kongsfjorden area, but it seems possible to us that it has become completely silicified. The Upper Gypsiferous Series deteriorates towards the west and north-west; we observed gypsum in the range of the Pallatiumfjella, but further north-west in Palasset the higher portion of the Cyathophyllum Limestone seems to be very porous and cellular without distinct gypsum. This coincides with the observations of Holtedahl (1913) in

the Dronningfjella where the upper part of the Cyathophyllum Limestone shows a similar development. This cellular limestone or dolomite has not been examined by us but it seems likely to be the equivalent of the evaporites. In Kongsfjorden itself the Cyathophyllum Limestone is uniform and only interrupted by one " Fusulina " limestone band which also occurs in the Dronningfjellet section of Holtedahl. This may correlate with the " Fusulina " limestone of the Ekmanfjorden area.

Palaeontological Age

No systematic collecting of fossils was carried out, and the following is an incomplete list of specimens obtained:

Limestone series: *Lonsdaleia* sp., *Fenestella* sp.

The following foraminifera have been identified from the fusulinid bed by Dr. R. Cummings: *Schubertella transitoria* (Staff and Wedekind), *Triticites cf. arctica* (Staff and Wedekind), *Triticites* spp.

The assemblage suggests an uppermost Carboniferous age.

Cherts: *Dictyoclostus* sp., *Neospirifer* sp., *Chonetes* sp. (?), *Tabulipora* sp., *Fenestella* sp., *Crinoid ossicles.*

No accurate determination of age can be made from this, but will have to await the results of work by Dr. C. L. Forbes on the Billefjorden faunas.

Acknowledgments

The expedition was made possible by a great number of contributions which will be acknowledged elsewhere. However, we would like to thank the Queen's University of Belfast and the Everest Foundation Trust Fund for their considerable financial help.

Our personal thanks are due to the Norsk Polarinstitutt for supply of survey data and aerial photographs; Col. D. R. Crone for his assistance in compiling the topographical map; Mr. W. B. Harland for his extensive advice to the expedition; Dr. R. Cummings for identification of foraminifera; Prof. A. Williams and Dr. J. Preston for reading the manuscript; and finally to the other members of the expedition.

REFERENCES

Doeglass, D. J., 1956. An Exponential Function of Size Frequency Distributions of Sediments. *Geol. en Mijnbouw.*, NW., SER., 18 Jaargang 1–29.

Føyn, S., and A. Heintz, 1943. The Downtonian and Devonian Vertebrates of Spitsbergen. The English-Norwegian-Swedish Expedition, 1939. *Skr. Norges Svalbard- og Ishavs-unders.* Oslo, Nr. 85.

Gee, E. R.,W. B. Harland, and J. R. H. McWhae, 1953. Geology of Central Vestspitsbergen: Part I Review of the Geology of Spitsbergen, with Special Reference to Central Vestspitsbergen;

Part II. Carboniferous to Lower Permian of Billefjorden. *Trans. Roy. Soc. Edin.*, lxii, 299–356.

HOLTEDAHL, O., 1912. Zur Kenntnis der Karbonablagerungen des Westlichen Spitzbergens. II. Allgemeine stratigraphische und tektonische Beobachtungen. *Vid.-Sesk. Skrifter. I. Math.-Naturv. KL.* Nr. 23, 1–91.

ISACHSEN, G., 1915. Travaux Topographiques de L'Expedition Isachsen 1909–1910. *Expedition Isachsen, Resultats Scientifiques*, ii, 1–61. Kristiania 1915.

MCWHAE, J. R. H., 1952. The Major Faultzone of Central Vestspitsbergen. *Quart. Journ. Geol. Soc.*, cviii, 209–230.

NORDENSKIOLD, A. E., 1876. Sketch of the Geology of Icesound and Bellsound, Spitsbergen. *Geol. Mag.* (II), III, 63–75.

NORSK POLARINSTITUTT. Chart 503, Svalbard, fra Bellsund til Forlandstrevet med Isfjorden. Oslo 1932.

ORVIN, A. K., 1934. Geology of the Kings Bay Region, Spitsbergen, with special reference to the Coal Deposits. *Skr. Norges Svalbard- og Ishavs-unders*, lvii, 1–195.

—— 1940. Outline of the Geological History of Spitsbergen. *Skr. Norges Svalbard- og Ishavs-unders.*, lxxviii, 1–57.

TYRRELL, G. W., and K. S. SANDFORD, 1933. Geology and Petrology of the Dolerites of Spitsbergen. *Proc. Roy. Soc. Edin.*, liii, 284–321.

DEPARTMENT OF GEOLOGY,
THE QUEEN'S UNIVERSITY,
BELFAST.

Sedimentary Dykes in the Dalradian of Scotland

By A. J. Smith and N. Rast

Abstract

Sedimentary dykes from the Dalradian of the South-west Highlands of Scotland are described. It is suggested that these features are characteristic of certain geosynclinal environments.

SEDIMENTARY, clastic, or sandstone dykes have been recorded and described from many parts of the world (see Diller 1890, Pavlow 1896, Arnold-Bemrose 1904, Vintage 1954). The structures described in the present communication are not the neptunian dykes in which material has accumulated particle by particle but those features in which forcible injection of sand has taken place.

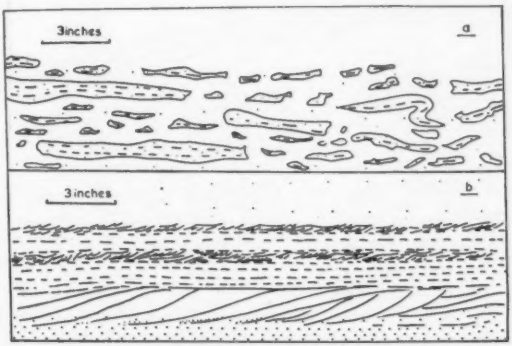
The most complete record obtained is from the Onich shore, Loch Leven, Argyllshire.

Onich Section

The best exposures occur on the shore of Loch Leven parallel to the road and between the tenth and the eleventh milestone from Fort William to Kinlochleven. The rocks occupy the eastern limb of the Appin Phyllite-Appin Limestone core of the Appin nappe (Bailey, 1910). At the western end of the section Appin Phyllites are predominant. Here the quartzose and pelitic beds are interbanded with fairly thin limestones and impure pebbly quartzites. Most of the eastern part of the section consists of the Appin Limestone, which locally is of a variety known as " tiger rock ", being an orange to yellow weathering sediment with numerous pale grey, white, and dark grey stripes in it. Normally these stripes are found to be lenticular and are occasionally contorted within the bed. There is no doubt that this contortion is of a primary sedimentary origin. The purer carbonate bands commonly contain large numbers of moderately rounded pebbles of quartz and feldspar which vary in size from 1 mm. to 25 mm. across. A few pebbles of felsite have also been recorded (Bailey, 1934, p. 512). The carbonate matrix varies from magnesian limestone to pure dolomite (Bailey, 1910, p. 596) weathering orange to pale bluish-grey, but being on fracture pale grey or sugary white. A large number of limestone bands can be recognized as discrete slump sheets (Text-fig. 1*a*).

At the junction with the Appin Phyllite and Limestone rhythmic deposits consisting of three phases (Text-fig. 1*b*) are encountered. The rhythm consists of (*a*) a pale grey coarse pebbly layer with large-scale current bedding (Text-fig. 1*b*) 3–6 inches thick, (*b*) a grey silty horizon separated from the other two phases by rapid transitions and at the top (*c*) a white or yellow-orange carbonate horizon about 3 inches

thick. The silty horizon often shows variations in thickness and is usually finely laminated. Both it and the succeeding carbonate layer commonly show small-scale ripple current bedding. In these beds, as well as in the Appin Limestone proper, rudimentary grading of pebbles is occasionally present. Such graded beds have the largest number of pebbles at the bottom, concentrated in a fairly well-defined layer above which the spatial frequency of coarse pebbles gradually decreases upwards. Nevertheless, the average size of the pebbles is constant throughout the bed. The pebbles with respect to their degree of roundness, size, and mineralogy are of the same type as those already described in the Appin Phyllite. Similar pebbles are present in the Appin Quartzite which stratigraphically underlies the Appin Limestone.



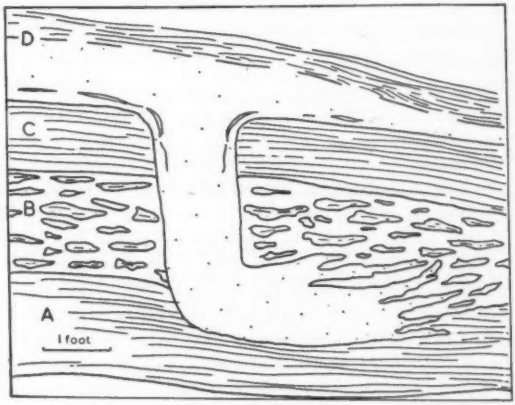
TEXT-FIG. 1*a*.—Slump sheet of pebbly limestone. The fragments consist of calcareous and gritty pelite.
b.—Example of a rhythm consisting of rudimentary graded and current bedded quartzite, pelite (dashed) and gritty carbonate deposit. (Openly stippled)—Ripple current bedded bands in the pelite and carbonate.

The carbonate or even the silty phase of the rhythmic deposits is overlain, not infrequently, by slump sheets which have an irregular lower surface with balls of pebbly material sinking in the underlying fine grain carbonate mud. The upper surfaces of such slump beds are quite distinct and are often succeeded by ripple current bedded carbonate-bearing silt. The slump sheets are composed of a carbonate-rich slurry which varies from an almost pure dolomitic limestone to a very impure carbonate. In this slurry there are frequent fragments of silty or even slaty material. Each fragment is usually drawn out into a fairly flat irregular lenticle, but some are contorted. The sheets vary in thickness from a few inches to several feet. Composite slumps are often present (cf. Wood and Smith, 1958). Slump folds are uncom-

mon throughout the section. Slumping is especially heavy in the Appin Limestone.

On the shore, below the prominent rampart of a bungalow situated 300 yards west of the eleventh milestone, the discrete slumps are involved in a sedimentary complex. Here a large number of such sheets have very little unslumped sediment between them, suggesting that the whole group was involved in slumping. It is this part of the section that shows many examples of sedimentary dykes. Two clear-cut varieties of dykes can be distinguished, namely the downward injected and the upward injected.

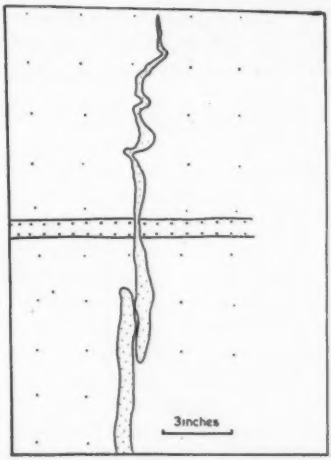
Downward Injected Dykes.—The features associated with the down-



TEXT-FIG. 2.—Downward injected dyke—description in the text.

ward injected dykes can be illustrated by reference to Text-fig. 2. This dyke originates at the lower surface of the bed D and is filled with unlaminated slurried material which forms the lower part of that bed. The dyke transgresses the laminated layer C and the slumped layer B, dispersing at the base of the latter in sill-like fashion into a series of branches. In the process a part of slightly laminated bed A has been removed. The ripple current bedding at the top part of the bed D leaves no doubt that the injection was downwards. While some fairly indistinct streaks at the bottom of the bed D are bent into parallelism with the margin of the dyke the laminated top of the bed is unaffected, indicating that the formation of the dyke took place before the topmost laminae of the bed D were deposited. In fact the formation of the dyke is practically contemporaneous with the deposition of the slurried bottom part of D. The fact that the dyke transgresses across the

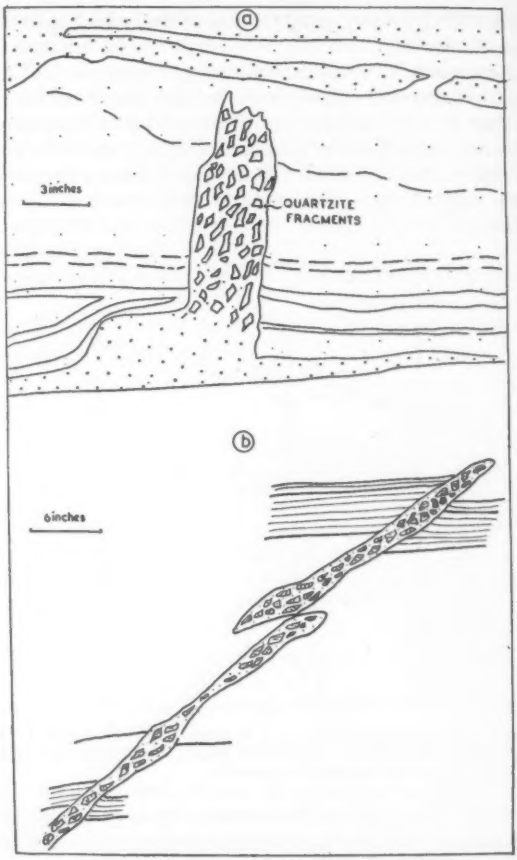
laminated bed C, but diverges at the level of the bed B, suggests that at the time of injection C was relatively brittle and rigid whereas B was yielding enough to accommodate the injected material. Bed B, being a slump sheet, probably still retained a considerable amount of trapped water at the time of the injection. The direction of branching of the sill is the same as the direction of inclination of fore-sets in the laminated and current bedded top of the bed D. Consequently the sill injected down the original slope. Downward injected dykes of this type are fairly rare.



TEXT-FIG. 3.—Homogeneous dyke of fine quartzite (finely stippled) injected upward into pebbly limestone (widely stippled) with bands of pebbly quartzite (coarsely stippled).

Upward Injected Dykes.—The upward injected dykes are quite common and vary in size from mere films to structures which are 1 ft. wide. In terms of the material which forms the dyke two sub-varieties can be recognized. In the first (Text-fig. 3) the infilling material is a fairly homogeneous psammite indistinguishable from the psammitic parent bed. In the second, however, there are a large number of angular often laminated fragments evidently derived from the parent bed and the walls of the dyke and injected together with finer matrix (Text-fig. 4*a* and 4*b*).

It is self-evident that the injection of both sub-varieties of the upward injected dykes, unlike the downward injected type, took place some time after the deposition of the parent bed. However, the uniform grain size in the first sub-variety implies that at the time of its



TEXT-FIG. 4 *a* and *b*.—Examples of upward injected dykes originating in consolidated quartzite (coarsely stippled) and injecting the pebbly limestone (finely stippled) and laminated pelite (horizontal lines). Thick continuous lines represent definite boundaries, while discontinuous lines are gradational junctions.

formation the matrix of the parent bed was entirely unconsolidated, whereas the breccia-like appearance of the dykes in Text-fig. 4 indicates a considerable degree of pre-injection consolidation.

The injection of all the sedimentary dykes evidently depends on the availability of fissures for penetration. Bailey and Weir (1932) suggest that the action of submarine earthquakes may provide such fissures.

Vintage (1954) proposes that faulting may be responsible for the formation of the fissures. In other words, tectonically unstable environments are implied by these workers. The sedimentary environment on the Onich shore again suggests instability. The presence of rudimentary graded bedding on one hand, and of fairly large-scale current bedding on the other, indicates that conditions in this area were intermediate between those of a typical geosynclinal trough receiving graded graywackes and a shallow shelf sea abounding in coarsely current bedded clastics and carbonates. Slumps and other features imply the existence of a slope on which periodic movements of sediments were initiated. The association of the dykes with the specially heavily slumped part of the section is suggestive of particularly unstable conditions which were possibly provoked by seismic activity.

Comparison with Perthshire–Loch Awe Succession

The sedimentary structures in the western and south-west Dalradians of Scotland have been hitherto described only as aids to the elucidation of complex tectonics. In recent years, encouraged by the publication of the work of Sutton and Watson (1954), the present authors have been systematically studying the Dalradian sedimentary structures from sedimentological point of view. Sedimentary dykes were found to be not very common except in rocks which in their other sedimentary features resemble the Onich shore formations. Particularly fine examples of sedimentary dykes are found in the Banded Group of the Blair Atholl Series in Perthshire and in the Tayvallich Limestone Series in Argyllshire. Many Perthshire exposures can be seen in the excellent outcrops of the Banded Group some three miles east of Kinloch Rannoch. In Argyllshire the western coast of the Tayvallich Peninsula also shows some fine examples. In these localities slumped sediments, the presence of both graded and fairly large-scale current bedding, and the association of medium to coarse clastics and limestones, again indicate the unstable conditions marginal to the geosynclinal troughs. The Tayvallich Limestone Series evidently represent such marginal conditions not only in Scotland, but also in Donegal where " veins of sand " have been observed by McCallien (1935).

B. A. Sturt (personal communication) had found sedimentary dykes to be a fairly common feature of the Carn Mairg Quartzite of Pitlochry area in Perthshire. The quartzite is not associated with limestones but nevertheless shows large-scale current bedding, rudimentary graded bedding, and slumps.

Conclusions

(1) In the Dalradian rocks of Scotland many localities show sedimentary injection dykes (cp. neptunian dykes). They are invariably

found within slumped rocks typical of the unstable marginal geosynclinal areas.

(2) Two varieties of injection dykes are recognized, namely, the downward injected contemporaneous with slumping and the upward injected which are clearly later than slumping.

Regarding the second conclusion Vintage (1954) reports that the downward injected sandstone dykes of Colorado possess features indicating a forcible injection of practically solid sand into fissures which were formed entirely later than deposition. For such structures Fairbridge (1946) suggests the term " cavity filling dykes ". Such a term would make the confusion with neptunian dykes inevitable and the present authors refrain from using it. On the other hand, in the same paper Fairbridge suggests the association of the sedimentary dykes with the slumped deposits of the marginal geosynclinal areas and this suggestion is confirmed by our observations.

REFERENCES

ARNOLD-BEMROSE, H. H., 1904. On some Quartzite Dykes in the Mountain-Limestone, near Snelston, Derbyshire. *Quart. Journ. Geol. Soc.*, xvi, 364–371.

BAILEY, E. B., 1910. Recumbent Folds in the Schists of the Scottish Highlands. *Quart. Journ. Geol. Soc.*, lxvi, 586–620.

—— 1934. West Highland Tectonics: Loch Leven to Glen Roy. *Quart. Journ. Geol. Soc.*, xc, 462–525.

—— and J. WEIR, 1932. Submarine Faulting in Kimmeridgian Times. East Sutherland. *Trans. Roy. Soc. Edin.*, lvii, 429–466.

DILLER, J. S., 1890. Sandstone Dikes. *Bull. Geol. Soc. Amer.*, i, 411–442.

FAIRBRIDGE, R. W., 1946. Submarine Slumping and Location of Oil bodies. *Bull. Amer. Assoc. Petr. Geol.*, xxx, 84–92.

MCCALLIEN, W. J., 1935. The Metamorphic Rocks of Inishowen. *Proc. Roy. Irish Acad.*, xlii, Sec. B, 407–442.

PAVLOW, A. P., 1896. On Dykes of Oligocene Sandstone in the Neocomian Clays of the district of Alatyr, in Russia. *Geol. Mag.*, iii, 49–53.

SUTTON, J., and J. WATSON, 1955. Deposition of the Upper Dalradian Rocks of Banffshire Coast. *Proc. Geol. Assoc.*, lxvi, 101–133.

WOOD, A., and A. J. SMITH, 1958. The Sedimentation and Sedimentary History of the Aberystwyth Grits (Upper Llandoverian). *Quart. Journ. Geol. Soc.* In press.

VINTAGE, P. W., 1954. Sandstone Dikes in the South Platte area, Colorado. *Journ. Geol.*, lxii, 493–500.

A.J.S.,
DEPARTMENT OF GEOLOGY,
UNIVERSITY COLLEGE,
LONDON.

N.R.,
DEPARTMENT OF GEOLOGY,
ABERYSTWYTH.

Ultrabasic Pillow Lavas from Cyprus

By I. G. Gass [1]

Abstract

Small isolated masses of highly olivine-rich lavas within a dominantly basaltic series of pillowed flows have been found in the northern foothills of the Troodos Igneous Massif. The rock is characterized by abundant phyric olivine usually set in a glassy, basaltic groundmass. Both intrusive and extrusive varieties of the rock have been found. It is suggested that the rock originated as a melt, possibly from the Troodos ultrabasic complex. Comparison is made with the serpentine lavas of Turkey.

I. Introduction

THE general geology of the Island of Cyprus is fairly well known. Previous workers, notably Bellamy and Jukes-Brown (1905), Cullis and Edge (1922), and Henson, Browne, and McGinty (1949), have shown that the island is divided into three main geological groups. The narrow east–west range to the north consists of thrust masses of Mesozoic rocks. The central, low-lying plain consists of Upper Cretaceous to Recent sediments, whilst the southern, Troodos range is made up of basic igneous rocks.

During the past five years parts of the Troodos range have been mapped in detail by members of the Cyprus Geological Survey under the direction first of Mr. D. W. Bishopp and, since 1953, Dr. F. T. Ingham. Work, mainly by Dr. Ingham, R. A. M. Wilson, J. M. Carr, and L. M. Bear, has resulted in the subdivision of the massif into three main groups. The central part of the range is occupied by an ultrabasic layered complex. This ultrabasic mass is surrounded by, and has been intruded into, the Diabase Formation, which consists of highly altered basic rock with a diagnostic, steeply dipping, sheeted structure. There is some difference of opinion as to whether the Diabase Formation represents a huge series of north–south dykes or a set of isoclinally folded lava flows.

Forming an incomplete ring around the Diabase is the Pillow Lava Series. This series has been subdivided by Wilson and Carr into Basal, Lower, and Upper divisions, mainly on the amount of intrusive material present. The Basal group is essentially similar to the Diabase Formation in that it consists of sheeted intrusives. Occasionally small pillow lava screens can be found and it is the occurrence of these extrusive rocks that identifies the division. The Lower Pillow Lavas contain abundant intrusives, both dykes and sills, but also have large areas of unintruded pillow lavas. The Upper division of the series has little intrusive material. It can be seen from the very nature of

[1] Published by permission of the Director, Geological Survey Department, Cyprus.

the criteria on which the subdivision is based that the boundary lines are usually transitional, vague, and very approximate. The only definite division is between the Lower and Upper pillow lavas. In this case Wilson and Carr have, in places, recognized an unconformity.

The sediments that overlie the pillow lavas range in age from Upper Cretaceous to Recent and are, for the main part, chalks and marls. Umber deposits are found in small lenses along the sedimentary-igneous contact. Associated with these umberiferous shales are cherts jaspers, and radiolarites. Radiolaria from one of these outcrops have been identified and dated as Jurassic, possibly Liassic, in age. To the south of the Troodos Massif, Upper Pillow Lavas are, in places, in contact with rocks belonging to the Mamonia Complex. Fossils from this complex have been dated as Triassic. The only contacts so far discovered where pillow lavas are overlain by the Mamonia Complex are possibly faulted. Sufficient proof of pre-Triassic age for the volcanic episode is not yet available.

It is not thought necessary in this paper to enter into a discussion as to the mode of origin of the pillow lavas. The thousands of feet of lava flows which comprise the volcanic series all show pillow structure. The view that these structures are formed by the extrusion of molten magma into a submarine or aqueous environment is accepted. Many of the phenomena put forward to support the submarine origin of pillow lavas (Schrock, 1948) are present in Cyprus. The pillows are well formed and usually have a marked chilled margin. Vesicles of varying size are universally present, and in many cases the interspaces between the individual pillows are infilled with calcareous, zeolitic, and, more rarely, umberiferous material.

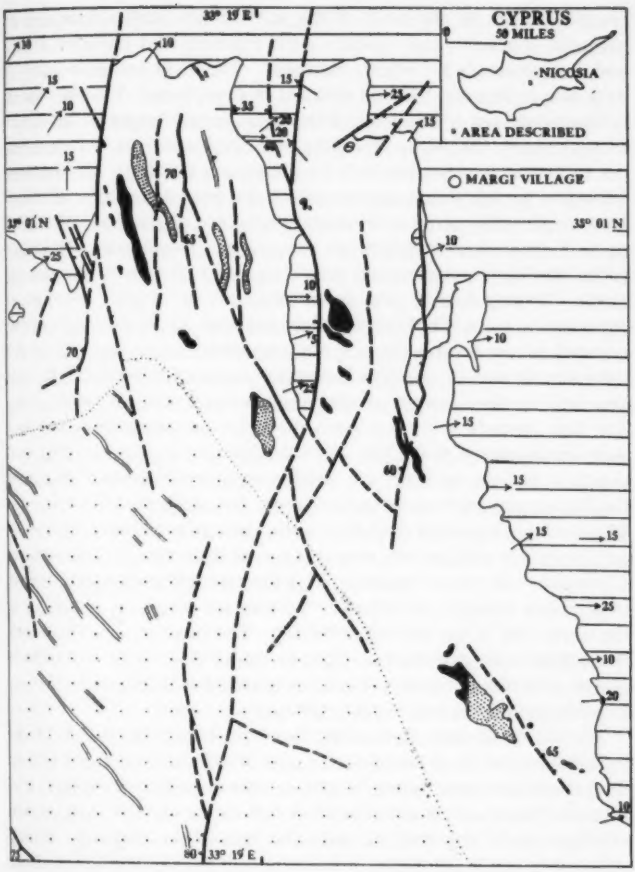
Small isolated outcrops of highly olivine-rich rocks have been located in other areas of the Troodos foothills. Although these ultrabasic rocks seem to be extremely rare, specimens having calcite pseudomorphs, most probably after olivine, are more common.

The ultrabasic rocks described here occur some 13 miles to the south-west of Nicosia, near the village of Margi and close to the igneous-sedimentary contact. The outcrops are isolated, varying in size from a few square feet to about 8,000 square yards; some consisting of entirely intrusive or entirely extrusive material. In other masses, both intrusive and extrusive types are represented.

II. Field Relations

The ultrabasic outcrops which occur within the area fall naturally into three groups (Text-fig. 1). One occurs to the west of the faulted sedimentary outlier, another to the east of this outlier, and the third group is in the south-east of the area shown on Text-fig. 1.

Of the seven large and the four small outcrops which form the



TEXT-FIG. 1.—Geological sketch-map showing the disposition of the ultrabasic rocks described.

group that lies to the west of the sedimentary outlier four are of extrusive character, one contains both extrusive and intrusive rocks, and the remainder are wholly intrusive. The main intrusive body is dyke-like in form and is well exposed in two places. The four small isolated outcrops to the west of the dyke are envisaged as offshoots from the main intrusion. The flow that emanated from the intrusion lies to its east and has been cut by subsequent faulting. All outcrops belonging to this group are formed of the vitrophyric type of ultrabasic rock, which consists of conspicuous green phenocrysts of olivine up to 5 centimetres in length set in a glassy groundmass. The dyke forms two large outcrops and possibly part of another further to the south. Poorly-defined columnar structure is visible throughout the intrusive mass. The contacts of the intrusive dip steeply and cut the adjacent pillow lavas. The five outcrops which lie to the east of the dyke are all part of one lava flow that emanated from this intrusion and they owe their present isolated nature to later north–south faulting. The flow extends 450 yards from the dyke and thins from 150 feet near the intrusion to 12 feet at its furthermost exposure. The two outcrops nearest the intrusive exhibit well-formed pillow structure, the pillows varying from 18 inches to 4ft. 6in. along the long diameter. They have well-marked chilled margins, the upper surfaces are convex and the lower surfaces are moulded round the underlying structures. Calcareous and zeolitic material, in places strongly ironstained, occurs interstitially between the pillows. Vesicles are common, especially in the upper half of the individual pillows. The two outcrops that mark the easternmost extremity of the ultrabasic flow show no definite pillow structure, but their extrusive nature has been confirmed by examination of top and bottom contacts.

The ultrabasic mass that occurs near and to the south-west of the sedimentary outlier, is regarded as a part of the group described above. In this outcrop both intrusive and extrusive rocks are exposed, the intrusive forms a thin north–south band on the western edge of the exposure whilst the flow occupies the rest of the outcrop. Pillow structure is better developed in this exposure than in any other of the ultrabasic extrusive masses. Well-defined cusps, selvedges, and chilled margins are present. The dip on all the extrusive members of the above group is to the east at between 30 and 60 degrees. The difference in degree of dip is probably due to tilting associated with the later faulting.

The second group of ultrabasic bodies which outcrops to the east of the sedimentary outlier is wholly intrusive. The ultrabasic masses stand out from the dominantly grey background of the basaltic pillow lavas. The largest mass is a plug forming a small but prominent black hill about 100 feet high and some 300 feet in diameter. Vertical

jointing is well defined. Associated with this main intrusive are a number of small sills and dykes that exhibit rough columnar jointing, a phenomenon common to other intrusives of the area. The ultrabasic rocks of this group are a holocrystalline type characterized by a slightly lower olivine content, extensive alteration of the olivine to serpentine, and a wholly crystalline groundmass. The plug is petrologically uniform.

In the south-east corner of the area shown on Text-fig. 1 there is a relatively large outcrop of ultrabasic rocks of the vitrophyric type. Typical intrusive material crops out to the west of the exposure whilst well-formed pillows are exposed in a small scarp to the east of the body.

To the east of the main outcrops described above there are a few small isolated masses of ultrabasic rock. These outcrops, usually less than 5 feet in diameter, are rounded in shape and probably represent part of the original flow which has been detached from its parent body by later flows.

The main ultrabasic outcrops are usually long narrow masses aligned in a north-north-west direction. This orientation conforms with the other lavas and intrusives in the area. No special tectonic forces are envisaged to aid in the extrusion or emplacement of the olivine-rich rocks. Their emplacement is not obviously associated with the extensive faulting that cuts the area.

III. Petrography

(*a*) *The "Normal" Pillow Lavas.*—Extensive alteration prevents, in many cases, accurate classification of the "normal" pillow lavas, but evidence so far collected shows that the bulk of the Upper Pillow Lavas are augite basalts. The centres of the pillows are usually so altered by propylitization, associated in many places with subsequent sulphide mineralization, that accurate optical determination of the plagioclase is rarely possible. In identifiable specimens it has been mainly determined as labradorite, but andesine and bytownite have been identified.

The bulk of the Pillow Lava Series is olivine-free. In some specimens calcite pseudomorphs after olivine are quite common. Usually the replacement of the olivine by the calcite is complete; occasionally only partial replacement has occurred. Calcite pseudomorphs have been found, forming between 5 and 35 per cent of the rock.

(*b*) *The Ultrabasic Rocks.*—The ultrabasic rocks within the area mapped are referred to as vitrophyric and holocrystalline.

Analyses and norms of the two rock types, together with an analysis of the olivine phenocrysts and calculated analyses of the groundmass of the vitrophyric type, are given in Table 1.

TABLE 1

ANALYSES

	(1)	(2)	(3)	(4)	(5)
SiO_2	43·00	42·11	42·4	44·75	51·81
TiO_2	0·18	0·23		0·49	0·57
Al_2O_3	4·64	6·59		12·52	14·50
Fe_2O_3	2·42	3·62		6·53	7·56
FeO	6·47	5·38	7·7	3·72	4·31
MnO	0·15	0·16		0·40	0·46
MgO	33·45	30·00	49·9	6·37	7·38
CaO	3·99	3·39		10·77	12·47
Na_2O	0·25	0·25		0·67	0·78
K_2O	0·05	0·17		0·13	0·15
H_2O^+	3·83	6·05		10·34	
H_2O^-	1·22	1·89		3·29	
P_2O_5	—	0·05			
Cr_2O_3	0·51	0·28			
NiO	—	0·12			
Total	100·16	100·29	100·00	99·98	99·99
S.G.	2·96	2·83	3·32		

NORMS

	(1)	(2)	(3)	(4)	(5)
q	—	—			13·8
or	0·3	1·1			0·9
ab	2·3	2·3			6·6
an	12·0	17·7			35·5
$CaSiO_3$	3·7	trace			11·0
$MgSiO_3$	24·3	30·6			18·5
$FeSiO_3$	3·0	2·8			1·7
Mg_2SiO_4	44·2	35·3			—
Fe_2SiO_4	5·3	3·5			—
mt	3·7	5·7			11·0
il	0·4	0·5			1·0
ap	—	0·1			—
cr	0·8	0·4			—

(1) Vitrophyric type of ultrabasic pillow lava, exposure to south-west of sedimentary outlier. Analyst: O. Von Knorring.
(2) Holocrystalline intrusive from plug to east of sedimentary outlier. Analyst: E. Waine.
(3) Olivine phenocrysts from same exposure as sp. 1. Analyst: Chr. G. Soteriades.
(4) Calculated analysis of the groundmass of the vitrophyric ultrabasic type. Percentages, by weight, taken in calculating groundmass analysis were olivine 62·3, chromite 1·0, groundmass 36·7.
(5) Calculated water free analysis of the groundmass, data as in (4).

Rocks belonging to this vitrophyric type are characterized by the abundance of fresh, euhedral phenocrysts of olivine, set in a glassy groundmass. The crystal outlines have been slightly modified by magmatic corrosion. The olivine phenocrysts whilst showing slight alteration to serpentine along fracture planes, are for the main part completely fresh. Small, usually rounded inclusions of the groundmass are commonly found within the phenocrysts. Throughout the many outcrops of this type there is very little variation in the ratio of olivine phenocrysts to groundmass. The average of six modes is olivine 56·9, groundmass 38·8, amygdales (calcite and chalcedony) 3·4, chromite 0·9. All modes of the ultrabasic rocks quoted in this paper are expressed as volume percentages and were determined optically by linear measurement of rock slices on an intergrating stage.

Optical examination of the olivine shows that it is colourless under ordinary light, has an imperfect cleavage, and is generally not zoned. Other optical data determined are tabulated below :—

$\alpha = 1{\cdot}646$, $\beta = 1{\cdot}665$, $\gamma = 1{\cdot}685$, $\gamma - \alpha = 0{\cdot}039 \pm 0{\cdot}007$, $(+)2V = 88°$.

All indices were determined in ordinary light by the immersion method and are correct to $\pm$ 0·004. Calculations made from this data give composition of Fa_7. The analysis of the olivine given in Table 1 confirms the optical data, giving a composition $Fa_{8.0}$.

The glassy groundmass is dark brown in colour, and has a refractive index of $1{\cdot}552 \pm {\cdot}004$, suggesting that it is basaltic in composition. Microlites of diopside are numerous in the groundmass. Much of the diopside has crystallized as long thin needles, but usually these are matted, in some cases the needles have a pseudo-spherulitic habit. On the crystal faces of the olivine there are often small diopside crystals that have grown at an angle to the surface of the phenocrysts. The only other primary mineral present is chromite. Vesicles are common and form about 3·5 per cent of the rock. The vesicles have been infilled with calcite and chalcedony, these minerals being present in roughly equal proportions. The size and shape of the cavities seemed to be determined by the disposition of the pyroxene microlites. The chalcedonic amygdales show the typical cryptocrystalline structure of that mineral. A rough annular structure is also visible, the more coarsely crystalline silica being in the centre of the vesicle.

The holocrystalline ultrabasic type differs from the vitrophyric type in that the groundmass is completely crystalline and the olivine has been more extensively altered to serpentine. The mode of this rock is olivine 45 (of which almost half is altered to antigorite), pyroxenes 21, anorthite 12 ; the remaining 22 is formed of secondary minerals,

mainly antigorite, zeolites, and magnetite. Some chromite is also present.

Generally, the size of the olivine phenocrysts tends to be slightly smaller than in the vitrophyric type; their outlines are euhedral, well defined, and clearly visible. The olivine has a biaxial positive sign, indicating a high Mg/Fe ratio. No zoning has been seen but the mineral is occasionally twinned. In most cases about 50 per cent of the olivine crystal has been altered to antigorite, but more rarely the alteration is complete. The antigorite adjacent to the olivine is straw yellow and has a lamellar habit whereas along fractures it is fibrous and grass green in colour. Throughout the rock there is abundant magnetite, often with skeletal structure. This mineral has evidently been separated during the alteration of the olivine.

The holocrystalline groundmass has a panidiomorphic texture and consists mainly of anorthite, diopside, and hypersthene. No olivine has been identified. Anorthite occurs as small subhedral laths, which in some cases show albite twinning. The maximum extinction angle, $\perp$ (010), on the laths showing albite twinning is 64 degrees, indicating a composition of Ab_5 An_{95}. Of the pyroxenes, diopside is by far the more abundant and forms typical squat prismatic crystals. It is colourless in thin section with (+) 2V about 60 degrees, $\gamma \wedge c = 37$ degrees. Twinning is common. The majority of the hypersthene crystals form as squat prisms but some have crystallized as long laths which tend to splay at the ends. Pleochroism displayed by this mineral is: X = pale pink, Y = very pale brown, Z = pale greenish-grey.

It is unusual, considering the altered state of the bulk of the pillow lavas, that many of the ultrabasic rocks should be so fresh. In the case of the vitrophyric type the glassy base has apparently protected the olivine phenocrysts from alteration. This suggestion is supported by the fact that in the holocrystalline variety the olivine phenocrysts have been extensively altered.

IV. Discussion

The analyses in Table 1 reveal that the rock owes its ultrabasic character essentially to the abundance of the olivine phenocrysts. The calculated analyses of the groundmass (columns 4 and 5, Table 1) indicate a basaltic composition. Table 2 below shows that the analyses of the Cyprus olivine-rich rocks compare very closely with those for peridotites. The comparison between the vitrophyric olivine-rich pillow lava from Cyprus (column 1), the average peridotite of Nockolds (column 3), and the peridotite from Harris (column 4) is particularly close. In hand specimen, however, the lavas could be defined as olivine-rich picrite basalts.

The presence in a nearby area of lavas containing varying amounts

of olivine, or more commonly calcite pseudomorphs after olivine, suggests the possible existence of a differentiation trend within the Pillow Lava Series. The rocks described in this paper may represent the ultrabasic end-product of such a differentiation process.

On the other hand, Muir, Tilley, and Scoon (1957) have shown that olivines from the picrite basalts of Kilauea range in composition

TABLE 2.—COMPARISON OF THE CYPRUS OLIVINE RICH ROCKS WITH OTHER ULTRABASIC ROCKS

	(1)	(2)	(3)	(4)	(5)	(6)	(7)
SiO_2	43·00	42·11	43·54	43·65	43·37	46·74	39·54
TiO_2	0·18	0·23	0·81	0·23	1·03	2·12	0·03
Al_2O_3	4·64	6·59	3·99	3·83	8·48	8·19	5·15
Fe_2O_3	2·42	3·62	2·51	2·24	2·91	2·77	3·48
FeO	6·47	5·38	9·84	8·61	11·00	9·6	4·39
MnO	0·15	0·16	0·21	0·17	0·13	0·14	0·14
MgO	33·45	30·00	34·02	33·10	25·93	20·47	32·71
CaO	3·99	3·39	3·46	3·44	5·03	7·47	4·06
Na_2O	0·25	0·25	0·56	0·64	1·33	1·74	0·14
K_2O	0·05	0·17	0·25	0·19	0·58	0·35	0·09
H_2O^+	3·83	6·05	0·76	2·26	0·19	—	8·81
H_2O^-	1·22	1·89	—	0·25	—	—	0·85
P_2O_5	—	0·05	0·05	0·06	—	0·22	0·13
Cr_2O_3	0·51	0·28	—	1·11	—	—	0·36
NiO	—	0·12	—	0·24	—	—	0·12
Rest	—	—	—	0·29	—	—	0·18
Total	100·16	100·29		100·31	100·25		100·18
S.G.	2·96	2·83	—	3·16	—	—	2·80

(1) Ultrabasic pillow lava : Cyprus. Vitrophyric type. Analyst : O. Von Knorring.
(2) Ultrabasic intrusive, Cyprus. Holocrystalline type. Analyst : E. Waine.
(3) Average Peridotite. Nockolds, 1954. *Bull. Geol. Soc. Amer.*, lxv, p. 1023.
(4) Peridotite, Harris, Inverness-shire. Analysis number 721, *Chem. Anal. of Ign. and Metamorphic rocks and minerals* 1931–1954. H.M. Stationery Office.
(5) Masafeurite. Johannsen, vol. iii, p. 332.
(6) Picrite-basalt, oceanite type ; average of nine analyses, Hawaii. G. A. Macdonald, *Hawaiian Petrographic Province*.
(7) Wehrlite, Troodos, Cyprus. Analyst : E. Waine.

from Fa_{14} to Fa_{20}. MacDonald (1949) has obtained similar values. Ross, Foster, and Myers (1954) have demonstrated that olivines from dunites and olivine-rich inclusions in basaltic rocks have a composition ranging from $Fa_{7.8}$ to $Fa_{11.0}$ (calculated from analyses). On this evidence the olivines from the Cyprus rocks, $Fa_{8.0}$, are akin to those from dunites and olivine inclusions in basaltic rocks rather than to the olivines of differentiation picrite basalts.

If, as the composition of the olivine suggests, these rocks are not

formed by crystal settling of olivine from a basaltic magma, two alternative modes of origin are possible. Firstly, the lava could represent a basalt containing a large percentage of olivine xenocrysts derived from the peridotite mantle, each xenolith being a single olivine phenocryst. Secondly, the ultrabasic lava could be formed from a melt of the same overall composition, derived either from complete fusion of part of the peridotite layer or from a peridotite body nearer the surface. In this connection it is significant that the lava described is homogeneous from the point of extrusion to the end of the flow, and that the phenocrysts are not interlocking or intergrown and do not tend to form aggregates. This would appear to support the second suggestion of crystallization from a peridotite melt.

The ultrabasic plutonic complex of Troodos, Wilson (1956), is thought to represent the welling up of the simatic layer through a profound fracture in the continental sialic crust. Wilson has mapped isolated offshoots of wehrlite emplaced into gabbroic rock some miles from the main ultrabasic mass. It is suggested that the olivine-rich lavas represent a further stage, where a melt from the peridotite zone of the Troodos ultrabasic complex was tapped off and found its way to the surface.

Sir E. B. Bailey and Professor W. J. McCallien (1953) have described the serpentine lavas of Turkey which, they consider, were originally " peridotite " lavas. They suggest that the original rocks were lavas consisting essentially of olivine, lubricated on extrusion by a trace of fluid material whose flow was assisted by the high pressure existing in a deep water environment. Nowhere did Bailey and McCallien find the unaltered rock from which their serpentine lavas were derived. It is believed that this unaltered rock could have been similar to the ultrabasic pillow lavas here described.

Acknowledgments

I wish to thank my colleagues of the Cyprus Geological Survey, especially Dr. F. T. Ingham, Mr. L. M. Bear, and Mr. R. A. M. Wilson, for their generous assistance during the preparation of this paper. My especial thanks are due to Dr. P. G. Harris, Lecturer in Geochemistry at Leeds University, for most helpful discussions on the petrological aspects, and to Dr. T. N. Clifford for his critical reading of the text.

REFERENCES

Bailey, Sir E. B., and W. J. McCallien, 1953. Serpentine Lavas, The Ankara Mélange and the Anatolian Thrust. *Trans. Roy. Soc. Edin.*, lxii, pt. ii.

Bellamy, C. V., and J. Jukes-Browne, 1905. *The Geology of Cyprus.* London.

Carr, J. M., and L. M. Bear, 1956. The Geology of the Peristerona-Lagoudhera Area. *Unpublished Cyprus Geological Survey Memoir.*

CULLIS, C. G., and A. B. EDGE, 1922. *Report on the Cupriferous Deposits of Cyprus*, London (Crown Agents for the Colonies).

HENSON, F. R. S., R. V. BROWNE, and J. MCGINTY, 1949. A Synopsis of the Stratigraphy and Geological History of Cyprus. *Quart. Journ. Geol. Soc.*, cv, 1–41.

MACDONALD, G. A., 1949. Hawaiian Petrographic Province. *Bull. Geol. Soc. Amer.*, lx, 1541–1596.

MUIR, I. D., and C. E. TILLEY, 1957. Contributions to the Geology of Hawaiian Basalts. The picrite Basalts of Kilauea : *Amer. Journ. Sci.*, cclv, 241–253.

ROSS, C. S., M. D. FOSTER, and A. T. MYERS, 1954. Origin of Dunites and Olivine-rich Inclusions in Basaltic Rocks. *Amer. Min.*, xxxix, 693–737.

SCHROCK, R. R., 1948. *Sequence in Layered Rocks.* McGraw-Hill, 359–368.

WILSON, R. A. M., 1956. The Geology of the Xeros–Troodos Area. *Unpublished Cyprus Geological Survey Memoir.*

GEOLOGICAL SURVEY DEPARTMENT,
NICOSIA,
CYPRUS.

The Use of Magnetic Measurements for the Study of the Structure of Talus Slopes

By Ian McDougall and Ronald Green

Abstract

By means of measurements of the direction of magnetization, talus may be distinguished from rock which has remained *in situ*. A criterion has been developed for the conditions under which the boundary between talus and rock *in situ* can be identified. This method is particularly useful when the upper portion of the rock *in situ* is broken and weathered and it has been applied with success to a scarp in the dolerite of the Western Tiers in Tasmania. Here the results suggest that in the upper part of the talus the jointed blocks have definitely fallen into a sub-horizontal position, whereas at lower levels they are only slightly tilted.

Introduction

It is of considerable economic importance in connection with tunnelling operations to know accurately the depth of talus covering undisturbed rock. In such cases the slope is usually drilled and an attempt is made to estimate the position of the junction between talus and rock *in situ* from core recovery, weathering, etc. In many cases it is extremely difficult to give an answer by drilling alone. One such case was encountered in the talus slope from a dolerite sill which had been drilled by the Hydro-Electric Commission of Tasmania and a preliminary study of this by Jaeger and Green (1958) showed that in this case magnetic measurements, and to some extent density measurements also, could be of considerable value in elucidating the structure of the material in the talus slope. It was found that rock *in situ* was magnetized in an almost vertical direction (i.e. a magnetic dip of almost 90°), whereas talus has very low values of magnetic dip. A detailed examination of the same talus slope has since been undertaken and in addition to an accurate and positive determination of the depth of talus cover it has been found that in the upper part of the talus the jointed blocks have definitely fallen into a sub-horizontal position whereas at lower levels they are only slightly tilted. This suggests that there are at least two stages in the formation of a talus slope; firstly, the columnar jointed blocks tilt, and subsequently the blocks fall over and come to rest in a sub-horizontal position.

While the case in Tasmania is extremely favourable for the application of the magnetic method, it is felt that the method can, in general, be of great value in elucidating the mechanics involved in the physical movement of rock masses. Aramaki and Akimoto (1957) have used an essentially similar method to estimate the temperature at which volcanic bombs landed.

The Geological Setting

The geology of the area has been mapped in detail by McKellar (1957) and only a short summary will be given here, the information being taken mainly from the above paper. Dolerite, probably of Jurassic age, was intruded in the form of a sheet some 1,200 feet thick into a sequence of gently dipping Permian and Triassic sediments whose total thickness was about 4,000 feet. The dolerite is confined almost entirely to the sediments of the Triassic system and in this region the roof sediments have been removed by erosion. During the Tertiary strong tensional faulting took place throughout Tasmania, and as a result of this the extensive Central Plateau was formed which on its eastern margin is bounded by a fault scarp with a relief of over 3,000 feet. The Western Tiers in this area were formed by the retreat of a scarp some three miles from the main north-west trending fault—the Tiers Fault. It is on this steep scarp that the talus deposits occur. The dolerite is broken up into a series of vertical contraction joint columns, commonly with sides up to 5 feet across, and less marked, well-spaced, approximately horizontal, joint planes. The petrology of the Great Lake dolerite sill has been dealt with by Jaeger and Joplin (1955), Joplin and Jaeger (1957), and McDougall (1958).

Results and Their Interpretation

The results of the measurements are tabulated in Table 1 and are graphically presented in Text-fig. 1. No account has been taken of the

Table 1.

Hole No. 5086.

Depth in feet below collar of hole.	*Inclination in Degrees.*	*Depth in feet below collar of hole.*	*Inclination in Degrees.*
150	86	750	86
300	88	801	89
400	86	850	84
497	88	900	86
701	90	1,000	86

Hole No. 5085.

Depth in feet below collar of hole.	*Inclination in Degrees.*	*Depth in feet below collar of hole.*	*Inclination in Degrees.*
4	65	172	54
16	20	176	70
27	62	180	86
39	37	190	87
50	26	201	80
80	36	225	87
100	31	251	80
112	41	275	85
125	66	306	87
150	59	400	79
160	61	450	90
170	61	510	86

Hole No. 5033.

Depth in feet below collar of hole.	Inclination in Degrees.	Depth in feet below collar of hole.	Inclination in Degrees.
25	5	260	4
37	4	270	2
49	7	280	52
65	20	283–300	15
76	22	300–314	84
85	35	317	90
99	9	350	89
113	12	373	90
124	30	400	82
138	28	426	88
151	15	455	85
164	14	459	90
174	18	474	88
201	22	502	89
214	34	526	82
225	44	548	90
251	17		

Hole No. 5032.

Depth in feet below collar of hole.	Inclination in Degrees.	Depth in feet below collar of hole.	Inclination in Degrees.
10	45	180	50
28	28	188	48
37	60	196	19
48	23	201	78
62	20	209	61
75	11	226	37
102	17	231	56
126	9	252	79
139	6	275	77
149	12	295	68
160	19	315	69
172	59		

Hole No. 5030.

Depth in feet below collar of hole.	Inclination in Degrees.	Depth in feet below collar of hole.	Inclination in Degrees.
50	28	100	69
75	11	110	81

Hole No. 5019.

Depth in feet below collar of hole.	Inclination in Degrees.	Depth in feet below collar of hole.	Inclination in Degrees.
12	15	199	59
31	42	226	68
48	50	249	58
75	56	276	63
99	50	300	70
125	57	329	56
152	58	363	65
175	58		

Hole No. 5020.

Depth in feet below collar of hole.	Inclination in Degrees.	Depth in feet below collar of hole.	Inclination in Degrees.
11	8	72	5
25	5	91	34
47	21		

The complement of the angle which the direction of magnetization makes with the direction of the borehole is listed for specimens from the given depths below the collar.

azimuth of the direction of magnetization because it is not possible to determine this from the borehole samples. In hole 5033 it can be seen that from the collar to a depth of 291 feet the samples have rather low dips which are almost without exception less than 45° and with the majority less than 35°.

The region from 283 feet to 314 feet is one of strong fracturing and very poor core recovery. From 314 feet to 328 feet there is still strong fracturing and very poor core recovery but the dip has now become high, and high values with an average of 87° are maintained to the base of the hole. This value of 87° is in close agreement with the measurements of Blackett (Jaeger and Joplin, 1955) on another borehole in the same igneous body, and with the mean value given by Irving (1956) for the whole of the Tasmanian dolerites.

These results suggest that all the dolerite below 314 feet (i.e. all the dolerite with a high dip) is *in situ* and that the transition to talus takes place in the region between 283 feet and 314 feet which consists of small, badly broken material. Above this region, judging by the fairly good core recovery, the material appears to consist of large blocks of elongated joint columns, which have broken off and whose low magnetic dips show that they are lying in a sub-horizontal position.

In holes 5019, 5020, and 5030 the relatively low dips indicate that all the material is talus (except possibly the lowest sample of 5030).

Borehole 5032 may possibly be an intermediate case. Here, low dips predominate to a depth of 196 feet, indicating that the material is entirely talus; below this depth the dips are steep, averaging 75°, but not so steep as that of dolerite *in situ*. This suggests that the lower part of this hole may be through joint columns which have tilted but not yet fallen over. The same appears to be true of hole 5085 since moderate dips occur to a depth of 176 feet, below which the dips become nearly 90° relative to the drill hole, but since the hole is at an angle of depression of 70° the dips in rock *in situ* would be expected to be of the order of 70°. However, there is also the possibility that even though the hole was commenced at an angle of depression of 70° it became vertical after a short distance owing to the drill being deflected by the very strong vertical jointing. This would explain very satisfactorily the high and uniform dips obtained below 176 feet, suggesting that below this level the dolerite is *in situ* and above it the rock is talus material.

These results imply that two distinct processes may be operative, firstly a tilting of blocks and secondly a definite falling over. The talus slopes on the face of the Western Tiers are in many cases not just groups of sub-horizontal columns with the complete range of azimuths but also there are very large blocks of dolerite—up to 300 yards across and 100 feet in height—which have moved as a unit and in which the

columns are dipping uniformly back in towards the dolerite cliffs behind. The relatively uniform dips of the magnetization for a number of successive specimens taken from a hole such as 5085 is most probably due to the fact that the hole passes through one of these large blocks.

Evidence from Density Measurements

Jaeger and Joplin (1955) and Jaeger and Green (1958) have shown that there is a characteristic variation in the density of dolerite as a sill is traversed from its base to its upper contact.

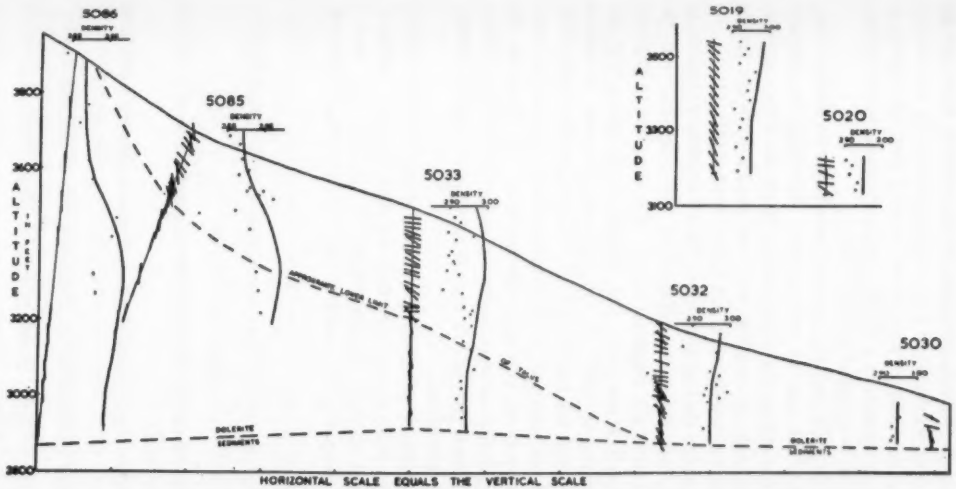
It was considered, therefore, that the height above the base of the sill at which a piece of dolerite was formed could be deduced from its density. Since the direction of movement of blocks in a talus slope is downhill, the recovery of a piece of core with a density characteristic of a higher level, indicates that the core has been obtained from a block which is talus. The densities of the samples from the bore cores under consideration were measured in the manner described by Jaeger and Green (1958) and the results plotted in Text-fig. 1. For comparison, a curve showing the expected variation of density with height above the lower contact for a bore through dolerite *in situ*, is drawn through the experimentally found points.

Statistical fluctuations are to be expected in the density determinations, but where there is a significant difference between experimentally found points and the curve for solid dolerite, it is to be expected that in this region the core has been passing through talus and not rock *in situ*.

From the collar of borehole 5033 to a depth of 300 feet, the significantly low densities indicate that in this region the borehole passes through talus and furthermore, that the talus has been derived from a considerably higher level in the sill. Below 300 feet the density values lie along the curve for solid dolerite, indicating that the borehole below this level is in rock *in situ*. For this borehole the density measurements are in excellent agreement with the magnetic measurements.

In 5085 the densities in the upper part of the hole agree with the results of a bore at the same level through dolerite *in situ*. The dolerite in the upper part of the hole is definitely talus material and appears from the density measurements to have originated from a level not far above its present position.

The density values in the lower part of the hole are less than is expected and this suggests that the dolerite was derived from a considerably higher level in the sill than that at which it now occurs. To reconcile this fact with the nearly vertical magnetization it is necessary to postulate some sort of landslide in which



TEXT-FIG. 1.—The results of density measurements and the direction of permanent magnetization of samples of core taken from seven drill holes into a talus slope on the Tasmanian Western Tiers are shown. The density determinations are shown by means of dots and the angle which the direction of permanent magnetization of each core sample makes with the drill hole is shown by short lines superimposed on the drill hole. The transition from low to high magnetic dips is taken to indicate the approximate lower limit of the talus.

a large block of dolerite broke off from the cliffs and moved down the slope some hundreds of feet with very little tilting during the movement.

In both 5019 and 5020 the densities of the samples show that the dolerite has a lower density than that to be expected if it were *in situ*, and indicate that the dolerite was derived from a position high up in the sill. This is in agreement with the magnetic results which indicate that the dolerite in both holes is not *in situ*.

The densities in 5030 and 5032 in the main conform closely to the values expected if the dolerite were *in situ*. Although the magnetic work indicates that at least the upper parts of the holes is talus material, this is not necessarily at variance with the results from the density measurements as the dolerite need not have travelled far.

Thus it can be seen that in the case of 5033 particularly, and also in 5019 and 5020, the density measurements are consistent with the results obtained from the magnetic work. The densities in 5030 and 5032 neither confirm nor deny the interpretation given from the magnetic results and in 5085 the densities would appear to indicate that the lower part of the hole is through a large dolerite block which, however, must have been part of a landslide which has moved several hundred feet downwards with but little rotation. Mechanically, this movement is of limited occurrence and the authors are more in favour of regarding the lower part of this hole as being *in situ* and the density results as being anomalous.

It is not within the scope of this paper to discuss the origin of these talus slopes other than to say that they are due to gravity acting on vertical elongated columns which, when left unsupported, tend to move down the steep slope and come to rest sub-horizontally at a lower surface.

The Conditions for the Identification of the Talus-Solid Rock Boundary

The following arbitrary directions are plotted on a stereographic projection (Text-fig. 2):—

(1) *M* the vector direction of the permanent magnetization of rock *in situ*.

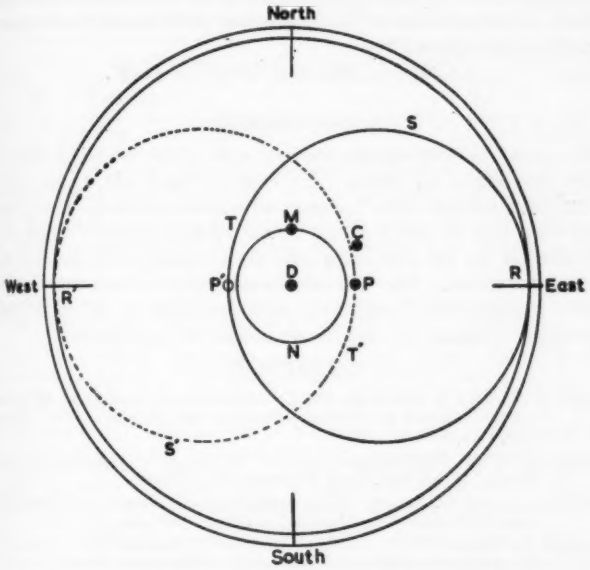
(2) *C* the vector direction of the elongated axis of the rock blocks.

(3) *P* the vector direction of the normal to the plane on to which the rock blocks fall.

(4) *D* the vector direction of the drill hole.

A system of co-ordinates is used in which the direction of *D* coincides with the OZ axis of the stereographic projection.

Writing $\angle$CM for the angle between the vectors C and M, this will also be the angle which the direction of magnetization in a fallen block makes with its elongated axis. However, the axes of the fallen blocks will be scattered with random directions over the plane whose normal



TEXT-FIG. 2.—A stereographic projection showing on the upper hemisphere the right circular cone MN along which must lie the directions of permanent magnetization of solid dolerite, and the right circular cones RST, on the upper hemisphere and R′S′T′, on the lower hemisphere, between which the direction of magnetization of talus material must lie.
These cones are only for the case discussed in the text.

is P. Thus the directions of magnetization found when drilling must be outside the right circular cones RST and R′S′T′ in the upper and lower hemispheres, respectively, whose axes are along P and whose semi-vertical angles are 90—$\angle$CM.

The direction of magnetization of rock *in situ* is given by the vector M, and as a result of the loss in azimuth when taking a bore core out of a drill hole the direction of magnetization for rock *in situ* may be anywhere on the circular cone MN.

The criteria for a positive determination between rock *in situ* and

talus is that the circles RST and MN should not intersect (this is the case shown in Text-fig. 2). This requires

$$\angle \mathrm{CM} + \angle \mathrm{MD} + \angle \mathrm{DP} < 90°.$$

For the Tasmaniam dolerites, *M*, *C*, and *D* are very close together and a sharp boundary can readily be identified.

Measurements of rock magnetism are of course treated statistically and if φ is the probable error in the direction of *M* the criteria developed above is to be replaced by

$$\angle \mathrm{CM} + \angle \mathrm{MD} + \angle \mathrm{DP} + 2\varphi < 90°.$$

ACKNOWLEDGMENTS

We gratefully acknowledge the help and advice we have received from Professor J. C. Jaeger. We wish to thank Mr. G. E. Hale, Geologist-in-Charge, H.E.C., for co-operation and help during the field work, and we are indebted to the Hydro-Electric Commission of Tasmania for the bore cores, and the University of Tasmania for transport facilities. The magnetic measurements were made on the astatic magnetometer constructed and described by Irving (1956), to whom our thanks are due for the use of the instrument.

REFERENCES

ARAMAKI, S., and S. AKIMOTO, 1957. Temperature Estimation of Pyroclastic Deposits by Natural Remanent Magnetism. *Amer. Journ. Sci.*, cclv, 619.

IRVING, E., 1956. The magnetization of the Mesozoic Dolerites of Tasmania. *Proc. Roy. Soc. Tasmania*, xc, 157–168.

JAEGER, J. C., and G. JOPLIN, 1955. Rock Magnetism and the Differentiation of Dolerite Sill. *Journ. Geol. Soc. Australia*, ii, 1–19.

—— and R. GREEN, 1958. A cross-section of a Tholeiite Sill. *Proc. of a Symposium on Dolerites.* University of Tasmania Press.

JOPLIN, G., and J. C. JAEGER, 1957. The Problem of the Quartz Dolerites: Some significant facts concerning mineral volume, grain size, and fabric. *Proc. Roy. Soc. Tasmania*, xci, 129–144.

MCDOUGALL, I., 1958. A note on the Petrology of the Great Lake North Sill. *Proc. of a Symposium on Dolerites.* University of Tasmania Press.

MCKELLAR, J. B. A., 1957. Geology of Portion of the Western Tiers. *Rec. of Queen Victoria Museum, Launceston, New Series*, vii.

DEPARTMENT OF GEOPHYSICS,
AUSTRALIAN NATIONAL UNIVERSITY,
CANBERRA.

CORRESPONDENCE

THE CARBONIFEROUS UPLAND FLORA

SIR,—Mr. R. Neves is to be congratulated on his paper (1958) on the spores of an Upper Carboniferous coal and its associated shales, which throws new light on a relatively unknown aspect of the Carboniferous flora. Mr. Neves shows that there is a striking abundance of pollen of the type produced by cordaites and conifers in his marine shale, in contrast to the associated coal and non-marine shale, where this pollen is a relatively trivial element. He also suggests that this is the case in other Carboniferous marine deposits. I believe that the implications of this are more far reaching than Mr. Neves suggests.

Spores in Carboniferous marine shale have a peculiar ecological significance in that they represent mainly the "spore rain" from what may be regarded as the permanent land of that time, in contradistinction to the spores of the swamp vegetation abundantly represented in coal. That cordaite and conifer pollen proves to be very abundant in such a marine shale is of particular interest in this context. Mr. Neves suggests that "a flora in which these plants (cordaites and conifers) abounded marginal to the marine depositional area could be envisaged as providing the high concentration of *Florinites* in the marine sediment". But the flora immediately adjacent to the marine depositional area would presumably be of the typical coal swamp type; and in this, as he shows from the coal itself, pollen of the cordaite and conifer type is a relatively trivial component.

A much simpler, and I believe more attractive, hypothesis is that the cordaite and conifer dominated vegetation was present throughout the depositional cycle in a situation practically unaffected by it—namely in the areas of higher land adjacent to the coal swamp. Before a marine invasion the spores of the lycopods, calamites, pteridosperms and ferns in the swamp in the central part of the basin would far outweigh locally the pollen rain coming from the relatively remote (if extensive) upland cordaite and conifer vegetation. When a marine inundation flooded the basin, the swamp vegetation would be temporarily displaced to a marginal belt, while that of the upland areas would be virtually unaffected. The marine mud deposited in the centre of the basin would continue to receive the same pollen rain from the uplands as before, but that from the restricted marginal swamp would now be in relatively lesser proportion. This hypothesis would explain Neves' observed abundance of *Florinites* as being a direct result of the marine phase.

An analagous situation to that postulated here occurs in the Tertiary pollen sequence in Venezuela (Kuyl *et al.*, 1955) in which changes in the pollen from the upland vegetation can be discerned against the background of locally dominant mangrove swamp vegetation which responds rapidly to minor changes in water level.

Whether this explanation of Neves' observations is valid or not, the spore contents of these marine shales are worth serious study, for these may be the only source of evidence that we shall ever have on the upland vegetation of that time. It is tempting to speculate that the true conifers, as distinct from the cordaites, may have been an important component of this upland flora during Carboniferous time long before they became a conspicuous element in the (macro-)fossil flora in the Permain. If we can learn to recognise the various Palaeozoic plants from their spores and pollen this type of palaeo-ecology holds interesting possibilities.

REFERENCES

KUYL, O. S., J. MULLER, and H. T., WATERBOLK, 1955. The application of palynology to oil geology with special reference to western Venezuela. *Geol. en Mijnbouw*, N. S. 17, 49–75.

NEVES, R. 1958. Upper Carboniferous Plant Spore Assemblages from the *Gastrioceras subcrenatum* horizon, North Staffordshire. *Geol. Mag.*, XCV, 1–19.

BOTANY DEPT., UNIVERSITY COLLEGE, LONDON.
12th April, 1958.

W. G. CHALONER.

NOTICE

UNIVERSITY OF NEW ENGLAND, ARMIDALE, N.S.W.

In February a fire at the University of New England destroyed a large percentage of the Department of Geology, including the entire Overseas Rock and Mineral collections, and the Collection of Departmental Reprints.

Whilst it has been possible to purchase selected overseas rock and mineral specimens, the bulk of the material destroyed was of a type not stocked by dealers, e.g. mineral and rock collections from particular areas possessing both teaching and research value.

The Department of Geology would be most grateful of any offers of duplicate reprints, rock and mineral collections that are of no further use to departments or individuals.

Inquiries should be addressed to : Professor A. H. Voisey, Department of Geology, The University of New England, Armidale, 5N, New South Wales.

PUBLICATIONS RECEIVED

BRITISH TERRITORIES IN BORNEO : Geological Survey Department. Mem. 8, *The Geology and Mineral Resources of the Upper Rajang and Adjacent Areas, Sarawak*. By H. J. C. KIRK. pp. xv + 181, with 29 figs., 30 tables, 53 pls., and 4 coloured maps in folder. Kuching, Sarawak, 1957. Price 21*s*.

BRITISH TERRITORIES IN BORNEO : *Annual Report of the Geological Survey for the year 1957*. By F. W. ROE. pp. x + 200, with 31 figs., 32 tables, and 43 pls. Kuching, Sarawak, 1958. Price 7*s*.

NIGERIA : *Geological Survey of Nigeria, Bull.* 27. The Geology of parts of Niger, Zaria, and Sokoto Provinces, by W. RUSS. 42 pp., with 13 plates (incl. 2 coloured maps), 2 figs., and 15 tables. Lagos, 1957, price 15*s*.

G. H. JONES. Memoria Explicativa y Mapa Geológico de la Región Oriental del Departamento de Canelones. *Instituto Geológico del Uruguay, Bol.* 34, 1956. 110 pp. with 42 pls. and folding map in colour (1 : 200,000).

REVIEWS

DIE GEOTEKTONISCHE ENTWICKLUNG DES APENNIN-SYSTEMS. By R. B. BEHRMANN. Geotektonische Forschungen, 12, pp. 1–99 with 15 figs. and 2 pls. Schweitzerbait'sche Verlag., Stuttgart, 1958. Price DM. 25·60.

This slim volume contains a very timely review of developing ideas on the structure and history of peninsular Italy and the Mediterranean basin in which it lies. The bibliography contains about 300 entries including a large number of recent publications as well as a valuable selection of earlier works. The substance of the paper is largely a review of many of the contributions so listed, but its form is far from an impersonal abstract. On the contrary, there is an ambitious attempt at a synthesis of that part of the Mediterranean region centred on the Apennines with special reference to its developmental stages.

Part I gives a brief palaeogeographical story right through the stratigraphical record. Part II, on the nature and scale of alpine tectonics in the Apennines and adjoining regions, plunges straight into the nappe structure of the northern area, to be followed in turn by similar outlines of the central and southern Apennines, the Calabrian Massif, Sicily, the Adriatic marginal trough, the Adriatic foreland, and the Tyrrhenian backland. Part III synthesizes still further and presents an overall picture of a developing geosyncline and orogenic belt along the length of the Italian peninsula and bounded by supposedly more stable areas in the sea to either side. The main Apennine folding, as is generally known, advanced from beyond the west coast progressively east-north-eastwards towards the Adriatic. With less area above sea level from which to interpret, a parallel and opposite movement of less extent is claimed for Corsica and Sicily, giving overall a somewhat symmetrical pattern and providing a schematic relationship with the Alps to the north and the Atlas to the south-west.

In such a brief compass discussion of detailed evidence could hardly be expected and indeed the whole treatment is essentially an interpretation. To the outsider without local knowledge this is an invaluable gateway to the literature of a region no less important than the Alps themselves. So often a simplified scheme may be conceived as a whole without undue effort and in so far as this is used provisionally to test observations rather than as a final solution it cannot fail to be beneficial. In making such use of it the reader is greatly assisted by the short paragraphs with key words emphasized and with liberal references as well as by the attractively diagrammatic maps and sections, and the author will not be to blame when some of the latter become detached from their context and used uncritically in still larger and more remote syntheses and as factual cross sections in general textbooks.

W. B. H.

STUDIES ON FOSSIL VERTEBRATES. Edited by T. S. WESTOLL. pp. xii + 263, with numerous text-figs. University of London, Athlone Press, 1958. Price 35*s*.

This volume consists of a collection of essays by some of the leading vertebrate palaeontologists, contributed and published in honour of D. M. S. Watson. The Editor in his preface expresses the regret of all concerned that the appearance of the volume could not more nearly coincide with Professor Watson's retirement from the Jodrell Chair of Zoology and Comparative Anatomy at University College, London.

There are in all eleven essays, covering a wide field of general and particular topics. Two of these, by A. Heintz and F. R. Parrington, are on the Anaspida, and Parrington adds an interesting section on mode of life of the Anaspids which provides a logical explanation of their posteriorly-situated gill openings. Three, by Sir Gavin de Beer and W. E. Swinton, J. Brough and W. K. Gregory, are essays on general evolutionary topics. T. S. Westoll, in addition

to editing the volume, contributes a chapter on the lateral fin-fold theory in relation to the pectoral fins of Ostracoderms and early fishes, and E. H. Colbert deals with tetrapod evolution in the Mesozoic under the title Beginning of the Age of Dinosaurs. E. S. Hills provides a general review of Australian fossil vertebrates, and A. S. Romer gives an account of the Permian Redbeds of Texas and their vertebrate faunas. E. I. White discusses the nature of the original environment of the Craniata, concluding that zoological, geological, and palaeogeographical evidence all point to an original off-shore distribution in coastal waters. Miss Rayner's contribution is the most geologically interesting, being an analysis of the sedimentational background of fossil fish occurrences (particularly in the Jurassic of Britain), and many readers will agree with her concluding remark that if the palaeontologist " wants to pursue problems of environment he will ignore sedimentary petrology at his peril ". The book is well produced and makes a fitting tribute to a famous palaeontologist.

O. M. B. B.

PHREATOPHYTES. By T. W. ROBINSON. United States Geological Survey Water-Supply Paper 1423. pp. 84, 32 figs. Washington, 1958. Price 40 cents.

This paper discusses the whole problem of the effect of certain plants (phreatophytes) which tap the ground water table or the capillary fringe in the desert areas of the Western United States. Most of these plants " have low economic value and consequently the water they use and return to the atmosphere without substantial benefit to man is defined as consumptive waste ". During the growth season in Arizona the Salt cedar (*Tamarix gallica*) has been shown to cause diurnal fluctuations in the water table of over 2 inches with a cumulative effect on stream flow. The annual use of water by such plants has been estimated to reach 7 acre-feet per acre.

Seventy plant species are classified as phreatophytes in the technical sense and eight of these have been submitted to detailed study.

Salvage of the ground water can be effected in two ways : " one by lowering the water-table by pumping beyond the reach of phreatophytes and subsequently using the water economically : and two by substituting plants of higher economic value." The latter would appear to offer greater benefit in the long run without risk of other difficulties consequent on pumping. These investigations must have considerable economic application in the many marginally desert areas of the world.

N. F. H.

FOSSILIUM CATALOGUS II : PLANTAE. Ed. W. JONGMANS. Part 33, Filicales, Pteridospermae, Cycadales, 6, by W. JONGMANS and S. J. DIJKSTRA. pp. 367–462. W. Junk, S'Gravenhage, April, 1958. Price fl. 30.

The staff of the Geological Bureau in Heerlen are to be congratulated on the continuation of this series so soon after the death of Professor Jongmans. Dr. S. J. Dijkstra, whose name now appears on the title page, is well known for his studies of Carboniferous and other plant microfossils from many parts of the world.

Part 33 carries the work to the conclusion of the species of *Callipteris*. As the series is likely to take many years to complete it would be helpful if each part bore a compilation closure date ; otherwise it may be necessary to await a final list of contents for this information. Merely to illustrate this point and not as criticism, it is perhaps worth drawing attention to the genus *Biscalitheca* Mammay, 1957 (March), which is omitted from Part 33. As in the previous parts, other information including Russian references is brought at least up to 1956.

N. F. H.